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Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
a Division of
The American Association of Petroleum Geologists

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Articles submitted for publication to the *Journal of Sedimentary Petrology* should have a prefatory abstract and sufficient postage should be sent by the author to pay the return of the manuscript as it has been found that many articles require more or less revision before publication.

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Surficial sediment in Lake Mead. By B. N. Rolfe.

Criteria used for distinguishing lacustrine and fluvial sediments in Tertiary beds of Uinta Basin, Utah. By M. Dane Picard.

Combined preferential staining and cellulose peel technique. by Harold J. Bissell.

Hydrothermal alteration of Phosphoria mudstones. By Lawrence F. Rooney.

The Petrology of the Whita Sandstone, southern Scotland. By A. E. M. Nairn.

Cyclic patterns in mechanically deposited Pennsylvanian limestones of northeastern Nevada. By R. H. Dott, Jr.

A new technique for measuring roundness of sand grains. By D. A. Robson.

The lithology, with particular reference to the cementation, of Upper Carboniferous sandstones in northern Derbyshire, England. By J. T. Greensmith.

A statistical study of heavy minerals in sands of the South River, Augusta County, Virginia. By Dorothy Carroll.

Size analysis of sand by a sedimentation technique. By D. M. Poole.

Evaporite facies. By L. I. Briggs.

Size distribution of zircon grains in some samples of the lower Beaumont Clay. By J. J. W. Rogers and W. F. Powell.

A proposed classification and terminology of rocks in the series bauxite-clay-iron oxide ore. By Jiří Konta.

The mechanical constitution of certain present-day Egyptian dune sands. By S. A. Harris.

Trends of sedimentation of Gondwana rocks of Giridih Basin, Bihar, India. By C. N. Rao.

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SEPTEMBER, 1957

PENNSYLVANIAN SANDSTONES OF THE EASTERN INTERIOR COAL BASIN[1]

RAYMOND SIEVER
Illinois Geological Survey, Urbana, Illinois[2]

ABSTRACT

The Pennsylvanian sandstones of the Eastern Interior Coal Basin were studied petrographically and stratigraphically to get a better understanding of source areas, sediment transport patterns, and conditions of sedimentation. The petrographic study was extended to adjoining basins, the Appalachian, Michigan, and northern Mid-Continent, to obtain a regional picture of mineral composition.

Stratigraphic patterns are described in terms of general character, lateral persistence, thickness, and contrasts between channel and sheet sandstones. The sheet sandstones are commonly thinner, finer grained, more argillaceous, and more widely persistent than channel sandstones. In the petrographic study, modal mineralogy, quartz variety, and rough size analyses were made of 176 samples and heavy minerals were estimated from 35 samples. The sandstones fall into three petrographic types that correlate with stratigraphic position: orthoquartzites in the basal zone (Caseyville Group), types intermediate between orthoquartzites and subgraywackes in the next higher zone (lower Tradewater Group) and subgraywackes in all higher groups. Although there are minor differences in composition between geographic provinces, the major aspect is one of homogeneity over the entire area of the north-central and northeastern states.

Current intensity, the primary factor in sedimentation conditions, varied greatly both locally and regionally, but the range of variability was more or less homogeneous throughout the stratigraphic section and in all of the geographic provinces. Conflicting evidence on marine or nonmarine origin of the sandstones is harmonized, to infer deposition in a combination of a variety of shallow water marine, deltaic, and coastal plain environments. Source areas and patterns of movement of detritus to sedimentary basins were largely inherited from early Pennsylvanian time. After Caseyville time, detritus from the east and northeast moved over the entire basin of deposition, vastly diluting material derived from the north and northwest. Continued mild erosion of source areas throughout Pennsylvanian time led to gradual but incomplete removal of a sedimentary cover and to the exposure of limited areas of igneous and metamorphic rocks.

INTRODUCTION

This report presents the results of a regional study of the sedimentary petrology and stratigraphy of the Pennsylvanian sandstones of the Eastern Interior Coal Basin, which includes a major part of the state of Illinois and adjacent parts of Indiana and western Kentucky. For a fuller understanding of some regional aspects of sedimentary petrology, the study was extended to include parts of nearby sedimentary basins—the Michigan Basin, the Appalachian Basin, and the northern Mid-Continent region (fig. 1).

The objective of the integrated study of sedimentary petrology and stratigraphy was a better appraisal of source areas, of the patterns of movement of clastic detritus from source to sedimentary basin, and of the physical factors that influenced sedimentation at the site of deposition. Such information leads to a more complete understanding of continental structure and the dynamics of late Paleozoic crustal movements. This investigation may also serve as an example of the utility of combining two somewhat complementary approaches to the study of sedimentary rocks—petrology and stratigraphy—for arriving at the objectives.

Primary emphasis was placed on the coarse clastics of the Eastern Interior Basin. It was the sandstones of this basin that were sampled in detail and with which the writer has had direct field and subsurface experience. But regional reconstructions of paleo-

[1] Published by permission of the Chief, Illinois State Geological Survey. A number of samples were contributed by Harold R. Wanless, University of Illinois, and Wallace Howe, Missouri Geological Survey. I was helped in the field at various times in the past several years by John W. Huddle, U. S. Geological Survey, Walter V. Searight and Wallace Howe, Missouri Geological Survey, Gilbert H. Cady and George M. Wilson, Illinois Geological Survey. Jack A. Simon and H. B. Willman, Illinois Geological Survey, and H. R. Wanless critically read the manuscript. Manuscript received August 3, 1956.

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geography and paleogeology cannot be completely based on the study of only one basin, and so perforce the sandstones of nearby basins were studied, primarily petrologically. Because the writer has had limited field experience in the nearby basins, many of the stratigraphic aspects discussed are taken from the works of others.

The study of one lithologic type, sandstone was undertaken because more interpretable information on patterns of clastic sedimentation can be gained from the petrology of sandstones than from other major lithologies, with the least expenditure of effort. General conclusions based on study of sandstones apply to the general picture of clastic deposition and are particularly appropriate for the study of the Pennsylvanian system of the north-central and northeastern United States, which is dominantly clastic throughout the whole section.

STRATIGRAPHIC PATTERNS

An excellent background of stratigraphic information is contained in a recent report by Wanless (1955). As Wanless (1955, p. 1795–1798) pointed out, the clastic ratio is relatively high in almost all parts of the Eastern Interior Coal Basin. In most parts of the basin, sandstone constitutes a significant, though highly variable, proportion of the clastic rocks. Total sandstone thicknesses range from less than 100 feet in western Illinois to more than 1200 feet in southern Illinois (Wanless, 1955, p. 1798). Both clastic ratio and sand-shale ratio increase from northwest to southeast across the basin, but this effect is partially due to a greater thickness of the almost completely clastic, very sandy, lower portions of the Pennsylvanian in the southeast. If the lower beds are subtracted, it becomes apparent that greatest clastic ratios and sand-shale ratios are found

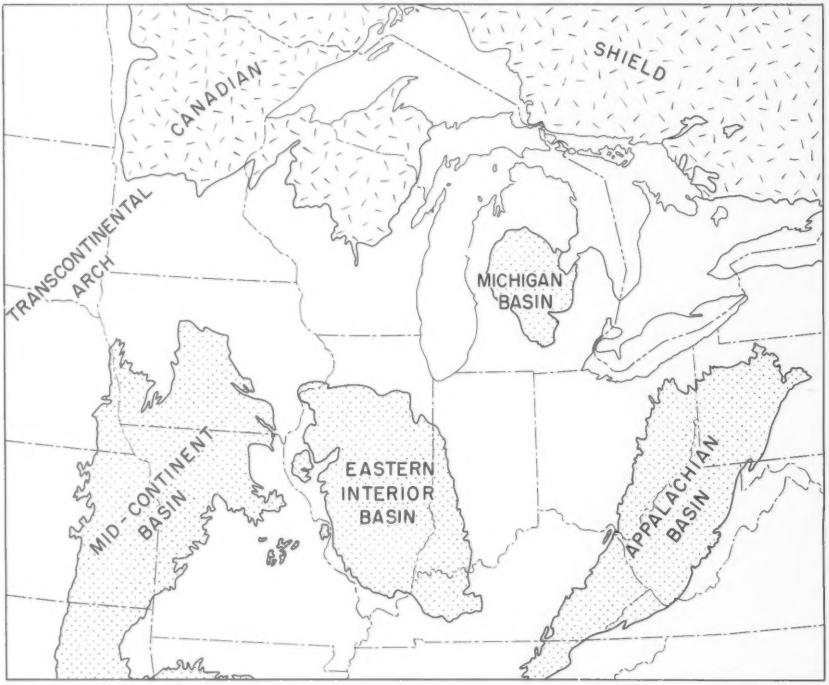

FIG. 1.—Index map showing area studied.

in the structurally deepest parts of the basin in southern Illinois. In general, minimum amounts of sandstone are found in the thinner sections characteristic of western Illinois.

The lowest stratigraphic group of the Pennsylvanian System in Illinois, the Caseyville, is composed largely of sandstones and conglomerates. Most of the pebbles in the conglomerates are quartz, but some fragments of shales, sandstones, and limestones are present in the lower portions of the fills of pre-Pennsylvanian channels (Siever, 1951, p. 577; Siever and Potter, 1956, p. 328). In most areas shale is the only other quantitatively significant lithology. In the lower half of the next higher group, the Tradewater, sandstones and shales are in about equal proportions with insignificant proportions of other lithologies. In the upper half of the Tradewater Group and in the higher Carbondale and McLeansboro Groups, the proportion of sandstone is smaller, shale is higher, and other lithologies, such as limestones and coal, assume significant proportions.

Two types of sandstone bodies in the Pennsylvanian of this area may be differentiated on the basis of shape—sheet and channel accumulations. A third type, shoestring sands, may be represented in relatively restricted areas along the LaSalle anticline in southeastern Illinois (Rich, 1916), but is unimportant quantitatively. There is no clear-cut division between many sheet and channel sandstones; the two types may be contemporaneous and grade into each other. Many sheet sandstones simply represent nonchannel phases, lying between areas in which channel sandstones developed. Other sheet sands, however, are fairly uniform in areal distribution and do not seem to be associated with any definite channel phases. The converse is not true—all channel sandstones are associated with thin sheet sandstones at one place or another.

Sheet sandstones are essentially absent in the Caseyville Group. In the lower half of the Tradewater Group in southern Illinois and western Kentucky some sheet sands are present but are minor in comparison with channel sands. In contrast, in western Illinois the Lower Tradewater sandstones are dominantly sheet accumulations, only the Bernadotte Sandstone occupying extensive channels. In the upper Tradewater, Carbondale, and McLeansboro Groups, sheet sandstones have a greater areal extent than channel sandstones in all parts of the basin. The channel sandstones in the higher groups are somewhat more important than the sheet sandstones in terms of total thickness and proportion of sandstone in the section.

The sheet sandstones are characteristically thin (5 to 25 feet), fine-grained, silty, and argillaceous. Bedding tends to be thin, and sedimentary structures are not abundant. Cross-bedding is uncommon, and rill marks, slump structures, and contemporaneous deformational structures are rare.[3] Both lower and upper contacts of many sheet sandstones are gradational although the upper contacts tend to be more gradational than the lower ones. Most of the sheet sandstones are conformable on underlying rocks.

The nature of the gradation from the underlying beds is revealed by many diamond drill cores. In many of the cores the underlying gray shale is interrupted, going upwards, by a few thin siltstone or very fine-grained sandstone laminae $\frac{1}{8}$ to $\frac{1}{4}$ inch thick. Higher in the core the laminae become thicker and more abundant until the rock grades to a fine-grained sandstone with thin shale interlaminations; still higher, the rock is a more or less uniform sandstone. Other sandstones show a relatively abrupt contact between the sandstone and underlying shale. Typically the upper contact of such sheet sandstones is a transition in grain size rather than interlaminations of shale and sandstone. In many sandstones the grain size decreases upwards, the rock becomes more argillaceous (interstitial rather than segregated into laminae) until the sandstone grades into a sandy or silty, poorly laminated shale or clay. However, this type of gradation is not everywhere present; transition zones of interlaminations as described for the lower contacts also occur at the tops of a few sandstones, and in rare instances the upper contact of the sandstone is abrupt.

The sheet sandstones tend to maintain their identity laterally and, even though

[3] See Potter and Glass (1957) for a detailed treatment of sedimentary structures in the Pennsylvanian rocks of an area in southern Illinois.

most do not have any distinctive lithologic features, they can be traced over moderately wide areas without great difficulty. Such lateral tracing is accomplished mainly by reference to key stratigraphic horizons above and below the sandstone, typically coal or limestone beds. Potter and Siever (1955) discussed the stratigraphic variability of the lower Pennsylvanian (Caseyville and Tradewater Groups) sandstones and came to the conclusion that the variability of the entire assemblage of sandstones in that stratigraphic interval was greater between township areas than within townships or within larger areas ("supertownships"). Qualitative observation by the writer suggests that the sheet sands show greater variability between the larger areas, such as counties, and less within such areas.

Channel sandstones are found in all parts of the Pennsylvanian column and in all areas of the Eastern Interior Coal Basin. This type of sandstone body is almost the exclusive one in the Caseyville Group and is dominant in the lower Tradewater Group of southern Illinois and western Kentucky. Higher in the Pennsylvanian the channel sandstones are less abundant.

The channel sandstones are quite variable in thickness, ranging from less than 5 feet at the borders of the channel to more than 150 feet near the middle of some channels. Many of the sandstones thicken from a few feet to well over 100 feet in a distance of one mile. Thickness is often related to an uneven erosional unconformity at the base of the sandstone.

Cross-bedding is one of the most prominent features of many of the channel sandstones. Although there has been no systematic measurement of cross-bedding direction on the sandstones of the entire basin, qualitative observation, plus the background of extensive measurement of cross-bedding in the Caseyville Group (Potter and Olson, 1954; Potter and Siever, 1956) suggest that the direction is fairly uniform in an outcrop and probably on a regional level. Observations made by Potter and Glass (1957) on middle and upper Pennsylvanian sandstones in southern Illinois and by the writer in western Illinois show that dominant directions of transport were to the south and southwest, indicating some persistence of the general direction of sediment transport from Caseyville time through the entire Pennsylvanian Period.

A variety of other sedimentary structures are found in the channel sandstones. Ripple marks, both current and oscillation, are abundant. Slump structures on both small and large scales are not uncommon. Other minor sedimentary structures such as groove casts and rill marks have been noted by Potter and Glass (1957).

Many of the channel sandstones lie unconformably on older beds. The thickness of older beds removed by erosion prior to sand deposition ranges from a few feet to 100 feet. As Weller (1930) noted, the bases of the erosional channels often rest upon coal beds. Weller implied from this that layers of compressed peat offered more resistance to erosion than beds of other lithologies. However, subsurface information in southern Illinois that has become available in the last 15 years indicates that many channels have cut through coal beds and that Weller's generalization has many exceptions.

The basal portions of many channel fills are conglomeratic. The pebbles in such conglomerates are largely of locally derived earlier Pennsylvanian sediments; shale, coal, and clay ironstone pebbles are the most abundant. The pebbles are generally poorly rounded and many have irregular shapes. The shape, roundness, and composition of these pebbles indicate relatively short distances of transport before deposition. Hopkins (1957), in a recent study, has shown the details of the morphology and composition of one prominent channel sandstone, the Anvil Rock, and its relationships with the associated sheet sandstone.

Extensive lateral variation characterizes almost all of the Pennsylvanian sands in the basin. Lateral variation includes two distinct but sometimes interrelated types, gradation in rock properties with little or no loss of the unit's identity as a bed or group of beds, and changes in identity of the unit resulting from thinning, wedge-outs, and abrupt facies changes. Gradation of rock properties is mainly variation in maximum and modal grain size and interstitial argillaceous content. In areas where such changes are extensive the identity of the unit also is affected.

An example of these kinds of variation and of the interrelationships between a channel and sheet sandstone is the Trivoli Sandstone (McLeansboro Group) of Williamson and Franklin counties in southern Illinois (Siever, 1949). Along the Williamson-Franklin county line in R. 3 and 4 E. the sandstone is in channel phase, reaching a maximum thickness of 157 feet. Through a belt 4 to 8 miles wide to the north of this area, the sandstone thins, loses its basal conglomerate, becomes slightly finer grained and less massive. Northward from this belt the sandstone becomes increasingly finer grained and argillaceous and interbedded with shale and shaly siltstone until in 3 or 4 miles it has lost its identity as a traceable unit. Farther north the sandstone reappears as a unit but as a thin (10 to 15 feet) sheet sandstone. Because the facies border between sandstone and silty shale is in the subsurface there is no firm indication as to whether it is erosional (silty shale older than sandstone) or sedimentational (silty shale contemporaneous with sandstone).

It is certainly true that many channel sandstones show sharp erosional lateral edges; it is equally true that a number of channels show gradational, interfingering edges. To make a tentative generalization, most of the channel sandstones of western Illinois show clearly from the sharp, abrupt bases and borders that the sands were deposited in erosional channels and many of the channel sandstones in southern Illinois show sedimentational rather than erosional facies changes.

In the highest part of the Pennsylvanian section, the upper part of the McLeansboro Group, a few sandstones grade vertically and possibly horizontally into limestones. For example, the Shumway Limestone in a part of Effingham County (NW $\frac{1}{4}$, SE $\frac{1}{4}$, Sec. 1, T. 6. N., R. 5. E.), in central Illinois, grades upward into a sandy limestone and then into a calcareous fine-grained sandstone. As far as is known, these occurrences are not common and are more or less restricted to the uppermost part of the system.

As mentioned above, sandstones are traced laterally chiefly by reference to overlying distinctive stratigraphic units. Figure 2, a cross section of Pennsylvanian strata through a part of southern Illinois, shows the efficiency of such tracing in the subsurface, by use of electric logs, rotary drill sample-study logs, and diamond drill core logs. The cross section illustrates the essential *small scale* lateral discontinuity of most of the sandstones and at the same time, the *large scale* persistence of many sandy zones. Even though any particular sandstone body may be interrupted in many places, nevertheless, the approximate zone is traceable over wide areas. Figure 3 shows that almost all sandstone beds can be traced if the datum points are sufficiently close together (in this case, oil wells 660 feet apart). Cross sections by Pryor (1956) show much the same features in some Pennsylvanian aquifers in White County, Illinois.

The Eastern Interior Coal Basin in Pennsylvanian time was composed of two structural elements: a rapidly subsiding basin area in southern Illinois, southwestern Indiana, and western Kentucky; and a slightly subsiding, more or less stable shelf or platform in western and northern Illinois (see isopach maps in Wanless, 1955, p. 1774–1779). There are no pronounced lithologic facies changes related to these structural elements, however. Most lithologies continue relatively unchanged from basin to shelf except for thickness changes that occur primarily in the sandstones and shales. Minor structural elements, such as several prominent anticlines in the structurally deep parts of the basin in southern Illinois, apparently influenced minor facies changes; these changes are primarily greater thicknesses of sandstone at the expense of shales over the high parts of the structures.

According to the theory of cyclical sedimentation of the Pennsylvanian in Illinois (Weller, 1930; Wanless, 1931; Wanless and Weller, 1932; Wanless, 1956), the use of the sandstone as the basal member of a stratigraphic unit, the cyclothem, was predicated on the fact that the sandstones were considered to be the earliest nonmarine beds deposited after an unconformity during which channels were cut subaerially. Information, both surface and subsurface, accumulated since the formulation of the theory, indicates that many sandstones are not unconformable on lower beds. The nonmarine origin of many of these sandstones also is open to question, as will be discussed later. These

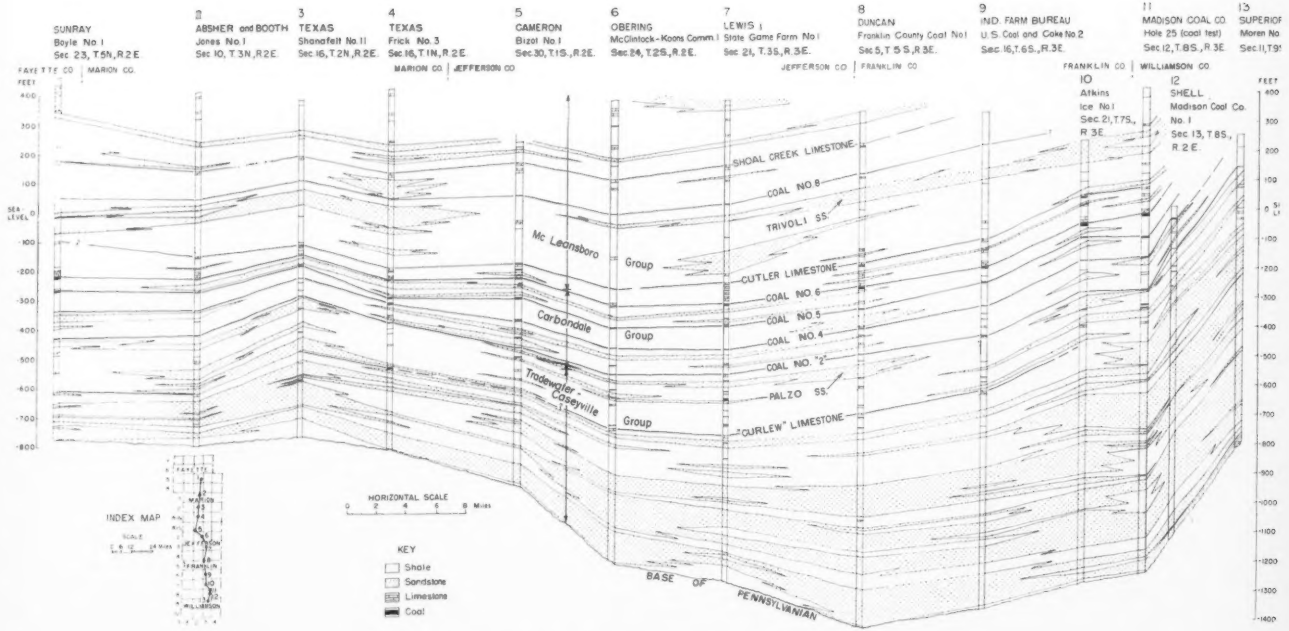

FIG. 2.—Cross section of Pennsylvanian strata in southern Illinois.

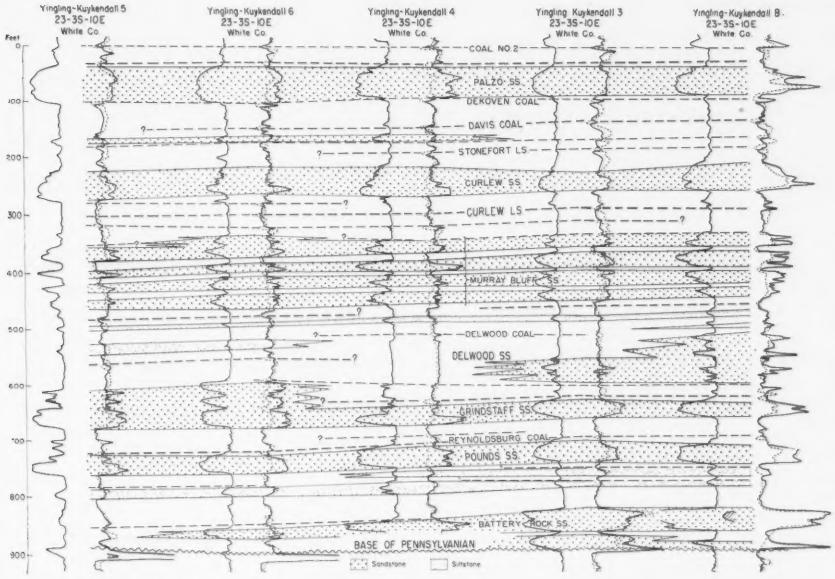

FIG. 3.—Cross section of Lower Pennsylvanian Strata in the Albion oil pool, Edwards County, Illinois.

factors modify the original theoretical conclusions that were used for justification of the classification scheme. The practical use of the sandstones as a classification boundary is limited by their high degree of lateral variation and discontinuity and the great difficulty of determining the position of the sandstone horizon in those subsurface sections in which the sandstone is absent.

SEDIMENTARY PETROLOGY

Relatively few sedimentary petrologic studies have been made of Pennsylvanian sandstones in the Eastern Interior Coal Basin. These include studies by Willman (1928), Gault (1939), MacVeigh (1932), Lamar and Grim (1937) Siever (1949), Berman (1953), Rusnak (1953), Biggs and Lamar (1955), and Siever and Potter (1956). Of these, only the last was an exhaustive study, including both thin-section and heavy mineral analyses of a large number of samples of the basal Pennsylvanian sandstones (primarily Caseyville Group). Studies of sandstones in the Mid-Continent include Clark (1950), Lins (1950), Mudge (1956), and Scruton (1950). Of these only Clark's and Scruton's are petrographic.

The sedimentary petrology of the basal sandstones was adequately covered by Siever and Potter (1956) and so only sandstones above the basal zone were sampled for this study. The samples were taken from sandstones throughout the entire Pennsylvanian section in many geographic areas in the north-central and northeastern states (fig. 4). Because concentration of effort was on the Eastern Interior Coal Basin, the largest number of samples was collected there. Complete coverage of all sandstones in the section was impossible but almost all sandstones of any great thickness or stratigraphic importance were sampled at least once (fig. 5). Many sandstones were sampled several times to get a better estimate of average composition and to appraise the general magnitude of variability of mineralogic composition and texture. Both diamond drill cores and outcrops of several sandstones were sampled to determine the effects of recent weathering on composition. Several sandstones were sampled at intervals

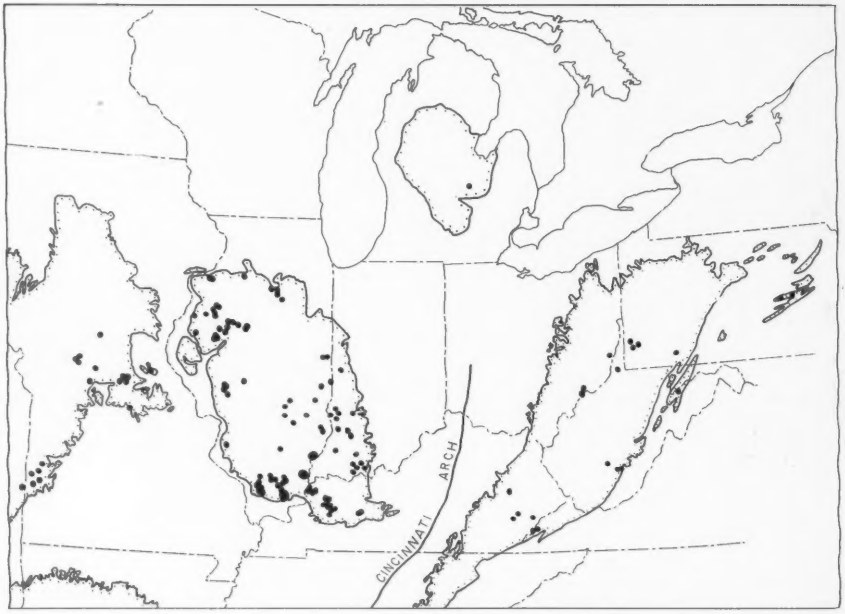

FIG. 4.—Location of petrographic samples.

from top to bottom to assess vertical variability within a sandstone at one spot. Multiple samples of sandstones within one geographic province were taken to evaluate lateral variability. A total of 134 samples were collected from the Eastern Interior Coal Basin.

The Pennsylvanian sandstones of the northern Mid-Continent, Michigan, and Appalachian Basins were sampled more at random. The samples from Missouri and Kansas extend over most of the stratigraphic interval but represent mainly stratigraphic rather than areal coverage. Only one sample was taken from the very small area and limited stratigraphic interval of Pennsylvanian rocks in the Michigan Basin. Samples from the Appalachian Basin cover the stratigraphic interval very unevenly. Because the Dunkard Series sandstones of southeastern Ohio, northern West Virginia, and southwestern Pennsylvania are essentially related to upper Pennsylvanian rocks, and indeed may prove to be partly upper Pennsylvanian in age (Cross, 1950), a suite of samples from this group of sandstones was analyzed for comparison with the Appalachian Basin samples. A few samples of sandstones and conglomerates from the western middle anthracite field of Pennsylvania were collected. A total of 42 samples were collected from basins other than the Eastern Interior.

The primary method of study was the petrographic analysis of thin sections. In addition, 35 heavy mineral separations from sandstones in the Eastern Interior Coal Basin were examined. Herbert D. Glass of the Illinois State Geological Survey provided clay mineral analyses of the <2 micron fraction from a few sandstones.

Modal analyses were made of all thin sections using a point counter and counting 200 points per slide. Only quartz, feldspar, clay (undifferentiated), carbonate minerals (including calcite, dolomite, and siderite), chert, rock fragments, muscovite, and other micaceous minerals (biotite and chlorite) were counted. Miscellaneous minerals such as hematite, limonite, and occasional heavy

minerals and pore space were not evaluated. Barite was included with carbonate minerals in those few sections in which it occurs. Percentages are given for total feldspar, but a detailed breakdown of proportions of plagioclase, orthoclase, and microcline is given in number of grains counted among the 200 grains used for the modal analysis. The only grains that were counted as rock fragments were those that showed definite detrital grain boundaries; equivocal grains that may have originated as clay floccules were included in the clay category. Grains counted as muscovite and other micaceous minerals were restricted to those of relatively large size that showed definite detrital boundaries and could not be considered in any way as part of the clay matrix.

Rough evaluations of size distribution were made from thin sections. They are given by maximum and minimum size (excluding clay matrix) and approximate modal size. Visual roundness of detrital quartz was estimated for modal sizes in most thin sections by use of the chart prepared by Krumbein (1941). No attempt was made to estimate roundness in some very fine-grained sandstones and siltstones or in sandstones in which original detrital outlines were obscured by corrosion, replacement, or secondary enlargement.

Quartz varieties were estimated by the same methods used by Siever and Potter (*1956*, p. *318*), counting 200 quartz grains per thin section and dividing the grains into three categories, igneous, metamorphic, and metamorphic quartzite. Heavy minerals were analyzed for relative abundance by visual estimation in terms of flood (>30%), abundant (>10%), present, and absent. Heavy minerals were separated from almost the entire sand fraction, .50 to .062 mm.

Modal, quartz variety, size and roundness, and heavy mineral analyses are shown in tables 1, 2, and 3.[4] The tables of analyses are arranged in stratigraphic order, oldest at bottom, and grouped by geographic province.

The modal analyses reveal a general pattern of successive sandstone types in most geographic provinces. The basal Caseyville Group and correlatives are orthoquartzites that contain, besides quartz, only minor amounts of clay, feldspar, and other constituents. (See Siever and Potter, 1956, for a complete discussion of this group.) The sandstones of the next higher interval, the lower part of the Tradewater Group and correlatives, fall between orthoquartzites and subgraywackes, containing more feldspar, clay, and mica than the Caseyville Group sandstones. The upper part of the Tradewater Group and all higher sandstones are subgraywackes that have high proportions of clay and significant amounts of feldspar, rock fragments, and mica. This sequence of sandstone types in one area in southern Illinois is discussed in detail by Potter and Glass (1957).

The boundaries between these sandstone types are not at all sharp. For example, in some parts of southern Illinois the Grindstaff Sandstone, just above the Caseyville Group, is a clean quartz sandstone that is almost indistinguishable from the orthoquartzitic quartz sandstones below. In many other areas, however, the Grindstaff (or equivalents) can be differentiated from the underlying sandstones by its clay, feldspar, and mica content. The upper part of the "transition zone" (term used by Potter and Glass, 1957, for the sandstones of the lower Tradewater Group) may contain sandstones that are petrographically similar to the higher subgraywackes.

The stratigraphic and physical boundaries between the three sandstone types are not only indistinct, but time transgressive as well. Thus the basal orthoquartzitic zone is younger in western and northern Illinois than it is in southern Illinois and western Kentucky, representing a transgression from south to north over the pre-Pennsylvanian unconformity. The boundary between the transition zone and the upper subgraywacke zone also transgresses time. In southern Illinois and western Kentucky the boundary is approximately at the position of the Curlew Sandstone (mid-Tradewater Group), whereas in western and northern Illinois the boundary is higher, at the level of the Isabel Sandstone (base of the Carbondale Group). In view of the relatively slow changes that are characteristic of intracratonic basins such as this, and the inevitable mixing of detritus from varied sources, it is doubtful whether the boundaries of these sandstone

[4] Table 2 shows locations of samples in table 1 by state and county. More precise locations are on file at the Illinois State Geological Survey.

TABLE 1.—*Modal analyses*

Sample number	Stratigraphic position	Quartz				Feldspar								Micaceous minerals		Size			
		Igneous	Metamorphic	Meta-quartzite	Total	Plagioclase	Orthoclase	Microcline	Total	Clay	Carbonate	Chert	Rock fragments	Muscovite	Biotite-Chlorite	Maximum	Minimum	Approx. mode	Average roundness
									ILLINOIS BASIN AND SOUTHERN ILLINOIS										
9	Upper Bogota	76.5	10.5	13.0	54.5	2	0	0	1.0	28.5	13.5	0	0	1.5	1.0	.12	.01	.06	—
305	Above Shumway Ls.	—	—	—	30.5	1	1	0	1.0	5.0	60.0	0.5	0	1.0	2.0	.10	.01	.05	—
258	Below "Watson" Coal	—	—	—	43.0	3	0	0	1.5	35.5	18.0	0	0	1.0	1.0	.12	.01	.05	—
307	"Divide"	85.0	7.0	8.0	50.5	4	1	0	2.5	8.0	33.5	0	0	3.0	2.5	.30	.01	.10	.4
330†	Mt. Carmel	—	—	—	64.0	—	—	—	5.5	24.5	0	0	4.0	2.0	0	—	—	—	—
301	Shoal Creek (Ss.)	72.5	12.5	15.0	65.5	6	2	0	4.0	30.0	0	0	0	0.5	0	.68	.02	.28	.5
326	Shoal Creek (Ss.)	76.5	7.0	16.5	59.0	2	2	0	2.0	32.0	4.0	0	0	2.5	0.5	.54	.02	.22	.5
328†	"Liberty"	—	—	—	59.0	—	—	—	8.5	26.0	0	1.5	1.0	4.0	0	—	—	—	—
329	Cutler	75.0	11.0	14.0	67.0	5	2	1	4.0	21.5	7.5	0	0	0	0	.56	.03	.24	.6
291†	Cuba	73.0	10.0	17.0	55.0	—	—	—	5.0	25.0	0	0	4.5	7.5	3.0	.52	.02	.20	.5
57	Palzo	69.0	17.0	14.0	67.5	6	4	4	7.0	20.5	0	1.0	2.0	0.5	1.5	.92	.04	.35	.6
292†	Palzo	73.0	11.0	16.0	66.0	—	—	—	4.0	25.0	0	0.5	2.5	1.5	0.5	.48	.02	.30	.6
261	Dekoven	74.0	14.5	11.5	59.5	3	3	0	3.0	33.5	0	0	0	3.5	0.5	.40	.01	.20	.4
58	Davis	79.0	9.0	12.0	56.0	7	2	0	4.5	35.0	0	0	0.5	3.5	0.5	.36	.02	.15	.5
59	Stonefort	76.5	11.0	12.5	69.5	7	3	1	5.5	22.5	0	0	0.5	2.0	0	.30	.02	.15	—
60	Lower Stonefort	76.0	10.0	14.0	66.0	5	5	1	5.5	25.5	0	0.5	0.5	0.5	1.5	.32	.02	.14	.5
61	Curlew	74.5	11.5	14.0	62.5	4	0	0	2.0	32.5	0	0.5	0.5	2.0	0	.42	.03	.22	.6
62	Curlew	71.0	11.5	17.5	70.5	5	3	0	4.0	22.5	0	0.5	0	1.0	1.5	.50	.02	.20	.5
303	Below Curlew	80.5	8.5	11.0	67.5	4	3	0	3.5	21.0	0.5	0	2.5	4.5	0.5	.40	.03	.16	.6
63	Murray Bluff	73.5	14.5	12.0	70.0	0	0	0	0	26.0	0	0	2.5	1.5	0	.82	.04	.36	.6
64	Delwood	78.5	9.5	12.0	67.0	0	1	0	0.5	28.0	0	0	1.0	3.5	0	.42	.04	.22	.5
65	Grindstaff	88.0	8.0	4.0	92.5	0	0	0	0	6.0	0	0	1.5	0	0	.42	.04	.26	.6
—*	Average of 4 Caseyville analyses	79.4	12.4	8.2	88.9	—	—	—	0.1	10.1	0	0.8	0.1	—	—	—	—	—	—
									New Haven core—White County, Ill.										
4-12	Trivoli	80.5	13.0	6.5	58.5	11	3	0	7.0	18.0	11.0	0.5	0	2.0	3.0	.30	.03	.15	.5
4-16	Below West Franklin Ls.	83.5	12.0	4.5	61.0	3	1	0	2.0	31.5	4.5	0.5	0	0.5	0	.12	.01	.06	.4
4-50	Anvil Rock	83.5	12.0	4.5	53.5	2	3	0	2.5	21.5	17.0	0	0	4.0	1.5	.20	.03	.10	—
4-60	Cuba	81.5	10.5	8.0	63.5	3	5	0	4.0	19.5	6.5	0	0	5.0	1.5	.32	.03	.15	.5
4-60C	Below Coal No. 5a	84.0	10.0	6.5	68.5	5	2	0	3.5	21.5	3.0	0.5	0.5	2.5	0	.30	.03	.12	.5
4-70	Below Coal No. 5	90.5	6.5	3.0	43.5	2	3	0	2.5	24.0	28.5	0.5	0	1.0	0	.25	.02	.09	.5
4-71	Below Coal No. 5	84.0	11.0	5.0	59.5	4	3	0	3.5	29.0	5.0	0.5	1.0	1.0	0.5	.20	.02	.10	—
4-87	Pleasantview	92.0	6.0	2.0	60.5	2	1	0	1.5	17.5	19.5	0	0	0.5	0.5	.30	.03	.08	—
4-97	Below Coal No. 2	88.5	8.5	3.0	76.0	0	2	0	1.0	14.5	5.5	0	0	2.5	0.5	.24	.02	.08	—
4-110	Palzo	90.0	5.0	5.0	50.0	3	0	3	3.0	10.0	35.0	0	0.5	1.5	0	.34	.02	.09	—
4-119	Dekoven	87.5	10.0	2.5	53.5	2	2	1	2.5	25.0	12.5	0	0	6.5	0	.37	.01	.09	—
4-127	Davis	87.5	6.5	6.0	78.5	2	5	0	3.5	11.0	3.0	0.5	0	3.0	0.5	.48	.03	.15	.6
4-140	Curlew	84.0	12.5	3.5	60.5	3	1	0	2.0	22.0	12.0	0	0.5	2.5	0.5	.38	.01	.10	.6
4-151	Murray Bluff	84.0	6.5	9.5	90.5	5	0	1	3.0	5.0	0	0	0	1.5	0	.64	.05	.30	.6
4-155	Murray Bluff	79.5	13.5	7.0	81.5	2	3	1	3.0	8.0	6.0	0.5	0.5	0.5	0	.80	.05	.40	.7
4-160	Murray Bluff	74.0	11.5	14.5	72.0	7	3	2	6.0	10.0	11.0	0	0.5	0.5	0	1.10	.04	.45	.7
4-168	Murray Bluff	80.5	10.5	9.0	78.0	4	4	4	6.0	11.0	3.5	0.5	0.5	0.5	0	.90	.05	.55	.7
4-182	Delwood	89.0	7.0	4.0	51.5	3	1	0	2.0	46.5	0	0	0	0	0	.44	.02	.13	.6
4-183	Delwood	84.0	9.0	7.0	59.5	3	0	0	1.5	2.5	35.0	0.5	0	1.0	0	.56	.03	.25	—
4-190	Delwood	87.5	9.0	3.5	54.5	0	1	1	2.0	33.0	10.0	0	0.5	0	0	.48	.02	.27	.6
4-202	Delwood	91.5	5.0	3.5	93.5	1	3	0	4.0	1.5	0	1.0	0	0	0	.64	.05	.25	.7
4-213	Above Willis Coal	92.5	6.0	1.5	69.0	0	0	0	0	31.0	0	0	0	0	0	.25	.02	.10	.5
4-219	Grindstaff	95.0	3.0	2.0	83.5	0	0	0	0	16.5	0	0	0	0	0	.24	.04	.15	.6
4-420A	Pounds	90.0	7.0	3.0	66.5	—	—	—	1.0	29.0	3.0	0	0.5	—	—	.34	.02	.10	.6
4-251	Battery Rock	91.5	6.0	2.5	27.5	0	0	0	0	2.0	24.0	0	1.5	—	—	.50	.02	.20	.7

Rotary drill cores—Edwards County, Illinois																			
C-1466	Mt. Carmel	84.0	11.0	5.0	67.5	10	2	2	7.0	17.5	5.5	1.0	0	0	1.5	.50	.02	.24	.6
C-1360B	Palzo	88.5	6.5	5.0	65.5	2	0	0	1.0	29.0	0	1.0	0	1.5	2.0	.36	.02	.14	.5
C-1360A	Curlew	83.5	8.0	8.5	71.5	5	1	2	4.0	21.5	2.0	0	0	1.0	0	.84	.01	.25	.6
C-1346B	Murray Bluff	92.5	6.0	1.5	83.5	1	2	0	1.5	14.0	0	0	0	1.0	0	.30	.03	.18	.6
C-1346A	Delwood	87.5	8.5	4.0	66.0	6	0	0	3.0	10.5	19.0	1.5	0	0	0	.64	.03	.30	.6
C-1465B	Grindstaff	94.5	5.0	0.5	76.0	1	0	0	0.5	6.5	16.5	0.5	0	0	0	.44	.03	.16	.7
C-1465A	Pounds	82.5	12.0	5.5	87.5	—	—	—	0.5	9.5	0.5	1.0	1.0	—	—	.35	.02	.16	.7
C-1319B	Battery Rock	92.0	6.0	2.0	97.0	0	0	0	0	2.5	0.5	0	0	—	—	.44	.04	.14	.7
Miscellaneous rotary drill cores																			
353	Caseyville	70.0	30.0	0.0	86.0	0	0	0	0	2.5	8.5	0.5	2.5	0	0	6.50	—	3.00	.9
C-1434	Caseyville	64.0	14.5	21.5	85.5	0	0	0	0	2.0	9.5	2.0	1.0	0	0	1.20	—	.80	.8
WESTERN ILLINOIS																			
251	Gimlet	91.5	5.5	3.0	55.0	4	5	0	4.5	32.5	5.0	0	0.5	1.5	1.0	.21	.02	.11	.5
252	Gimlet	90.5	5.0	4.5	50.0	2	2	0	2.0	9.5	35.0	0	0	1.5	2.0	.18	.02	.11	.5
320	Gimlet	85.0	7.5	7.5	59.0	1	3	2	3.0	32.0	0	2.5	0	2.5	1.0	.27	.03	.14	.5
249	Copperas Creek	91.5	5.0	3.5	63.0	4	1	3	4.0	23.0	4.0	0.5	0.5	2.5	2.5	.23	.04	.11	.4
306	Copperas Creek	88.0	8.0	4.0	61.5	4	4	0	4.0	28.5	0	0	0.5	3.5	2.0	.25	.04	.16	.5
250	Cuba	92.5	6.0	1.5	55.0	3	4	0	3.5	31.5	8.5	0	0	1.0	0.5	.23	.03	.09	.4
321	Cuba	90.0	2.5	7.5	62.0	2	1	0	1.5	34.5	0	0.5	0.5	0.5	0.5	.32	.03	.17	.5
164	Pleasantview	87.0	8.0	5.0	58.0	2	2	2	3.0	34.0	0	1.5	0.5	2.0	1.0	.23	.03	.13	.5
246	Pleasantview	89.0	9.0	2.0	58.0	4	4	0	4.0	29.5	7.0	0.5	0.5	0.5	0	.23	.02	.11	.5
247	Pleasantview	85.5	11.5	3.0	58.0	4	4	0	4.0	35.5	0	1.0	1.0	0.5	0	.23	.02	.10	.4
248	Pleasantview	94.0	4.5	1.5	53.5	3	3	0	3.0	39.0	0.5	1.0	1.0	1.5	0.5	.28	.02	.10	.4
253	Pleasantview	92.5	6.5	1.0	54.5	2	3	0	2.5	27.5	13.5	0.5	0	1.0	0.5	.16	.01	.06	—
83	Isabel	80.0	12.0	8.0	63.5	0	2	1	1.5	28.5	0	0.5	0	2.5	3.5	.18	.01	.08	—
84	Isabel	78.0	13.0	9.0	63.0	3	3	0	3.0	29.0	0	0.5	0.5	2.0	2.0	.28	.02	.10	.5
158	Isabel	69.5	21.0	9.5	61.6	6	4	3	6.5	21.0	0	4.5	6.0	1.0	0	.86	.03	.20	.5
163	Greenbush	88.0	7.0	5.0	74.0	2	1	0	1.5	18.0	5.0	1.0	0	0.5	0	.42	.03	.09	.5
245	Middle Delong	95.0	4.0	1.0	77.5	0	1	0	0.5	21.5	0	0	0	0.5	0	.20	.02	.10	.4
85	Bernadotte	92.0	7.0	1.0	83.5	0	1	0	0.5	15.0	0	0	0	1.0	0	.20	.02	.07	.5
109	Bernadotte	91.0	7.5	1.5	75.5	0	0	0	0	24.0	0	0.5	0	0	0	.42	.01	.08	—
157	Bernadotte	90.0	8.0	2.0	70.0	5	3	0	4.0	23.0	0	0.5	1.0	1.0	0.5	.54	.02	.25	.6
108	Pope Creek	93.0	4.0	3.0	90.0	0	0	0	0	8.5	0	1.5	0	0	0	.28	.03	.10	—
82	Tarter	88.0	8.0	4.0	87.0	0	0	0	0	12.0	0	0	0	1.0	0	.24	.02	.11	.6
—*	Average of 5 Babylon samples	94.0	3.9	2.1	93.1	—	—	—	0.2	5.9	0	0.4	0.3	0.1	0	—	—	—	.8
EASTERN ILLINOIS																			
323	Below "Divide" Coal	71.0	14.5	14.5	55.5	16	5	2	11.5	26.0	0	0	1.0	3.0	3.0	.32	.03	.10	—
297	Under Livingston Ls.	87.0	7.5	5.5	32.5	2	1	0	1.5	16.0	41.5	0	0	2.5	6.0	.20	.01	.05	.4
327	Below "Fithian" Coal	76.5	14.0	9.5	47.5	7	2	0	4.5	42.0	0	0.5	0.5	4.0	1.0	.36	.02	.10	—
309	Over "Macoupin" Coal	69.5	12.5	18.0	61.5	5	3	0	4.0	28.0	0	0.5	1.5	2.0	2.5	.26	.02	.12	—
259	Above Lonsdale Ls.	83.0	7.0	10.0	46.0	5	3	0	4.0	30.5	15.5	0.5	1.0	1.5	1.0	.30	.02	.10	.5
260	Cuba (?)	64.0	15.0	21.0	56.5	7	4	0	5.5	30.0	0	2.5	5.5	0	0	.72	.03	.35	.6
NORTHERN ILLINOIS																			
325	Gimlet	73.0	12.0	15.0	58.5	5	2	0	3.5	35.5	0	0.5	1.0	0.5	0.5	.40	.02	.15	.5
304	Copperas Creek	75.5	8.5	16.0	52.0	8	3	0	5.5	18.0	22.0	0	0	2.0	0.5	.30	.02	.12	.5
308	Vermilionville	70.0	12.5	17.5	60.0	6	6	1	6.5	28.5	3.0	0.5	1.0	0.5	0	.44	.02	.16	.5
322	Vermilionville	77.5	9.0	13.5	65.4	5	1	3	4.5	27.0	2.0	0	0.5	1.5	0	.50	.02	.20	.5
—*	Average of 5 Tradewater Group samples	68.3	14.7	17.0	56.8	1.8	1.8	2.4	3.0	12.7	24.0	1.3	2.2	—	—	—	—	—	—
SOUTHWESTERN ILLINOIS																			
254	Carlinville	78.5	8.0	13.5	42.0	4	2	0	3.0	14.0	36.5	0	2.0	1.5	1.0	.28	.02	.07	.4
255	Trivoli	78.5	9.0	12.5	62.5	4	3	1	4.0	30.0	0	0	1.5	2.0	0	.24	.02	.08	.4
256	Below Scottville Coal	80.0	6.5	13.5	52.5	8	5	1	7.0	14.5	20.0	2.0	3.0	0	1.0	.30	.02	.12	.4
257	Above Bankston Fork Ls.	68.0	12.0	20.0	66.5	9	1	4	7.0	12.0	8.0	1.0	4.5	1.0	0	.46	.04	.20	.5
310	Above No. 4 Coal	80.5	9.0	10.5	64.0	7	4	0	5.5	27.5	0	0.5	0.5	1.5	0.5	.40	.02	.15	.5
206	Below No. 2 Coal	—	—	—	70.0	2	2	0	2.0	26.5	0	0	0	1.5	0	.10	.01	.04	—
324	Seahorne	81.5	8.0	10.5	78.5	2	2	0	2.0	17.5	0	0	1.5	0.5	0	.50	.03	.20	.5
76	Upper Makanda	80.5	7.0	12.5	89.0	1	3	1	2.5	4.5	3.5	0	0	0.5	0	.52	.03	.25	.6
75	Upper Makanda	77.0	8.5	14.5	68.0	5	1	0	3.0	13.0	15.5	0.5	0	0	0	.84	.04	.30	.6
319	Delwood	84.0	6.5	9.5	77.5	3	1	0	2.0	19.0	0	0	1.5	0	0	.44	.02	.16	.5
74	Boskydell	68.0	15.5	16.5	63.5	2	3	2	3.5	13.5	19.0	0	0.5	0	0	1.10	.06	.35	.6
—*	Average of 11 Caseyville samples	84.2	10.8	5.0	94.0	—	—	—	0.2	5.0	0	.7	.1	—	—	—	—	—	—

TABLE 1—(*continued*)

Sample number	Stratigraphic position	Quartz				Feldspar				Clay	Carbonate	Chert	Rock fragments	Micaceous minerals		Size			Average roundness
		Igneous	Metamorphic	Meta-quartzite	Total	Plagioclase	Orthoclase	Microcline	Total					Muscovite	Biotite-Chlorite	Maximum	Minimum	Approx. mode	
								WESTERN KENTUCKY											
235	Mt. Gilead	86.5	5.5	8.0	66.5	5	3	0	4.0	22.5	0	1.5	0	4.0	1.5	.32	.03	.10	—
233	Vanderburg	86.0	5.5	8.5	81.0	4	0	1	2.5	10.5	0	1.5	1.0	3.5	0	.40	.03	.14	.6
230	Dixon	82.5	6.5	11.0	62.0	1	4	0	2.5	33.5	0	0	0	1.5	0.5	.42	.03	.14	.5
234	Dixon	82.0	8.5	9.5	73.5	3	3	0	3.0	20.0	0	1.5	0.5	1.5	0	.34	.02	.14	.5
231	Above Madisonville Ls.	84.5	3.5	12.0	74.5	6	3	0	4.5	15.5	5.0	0	0	0.5	0	1.00	.05	.32	.5
269	Above No. 14 Coal	79.0	12.0	9.0	67.0	7	9	3	9.5	22.0	0	0	0	1.5	0	.86	.04	.40	.5
262	Anvil Rock	91.0	6.5	2.5	82.5	0	4	0	2.0	14.5	0	1.0	0	0	0	.48	.02	.22	.5
268	Below No. 11 Coal	94.5	3.0	2.5	60.5	2	4	0	3.0	32.5	0.5	0	0.5	3.0	0	.32	.02	.08	—
266	Above No. 9 Coal	85.0	10.0	5.0	56.5	1	3	0	2.0	40.5	0	0	0	1.0	0	.48	.02	.20	.5
263	Below No. 9 Coal	88.5	7.5	4.0	70.0	1	1	0	1.0	15.0	14.0	0	0	0	0	.64	.03	.26	.6
264	Above No. 8 Coal	95.0	2.5	2.5	69.5	0	0	0	0	18.0	12.0	0	0	0.5	0	.30	.02	.10	—
267	Below No. 8 Coal	88.0	7.0	5.0	75.5	1	4	0	2.5	19.0	0	1.5	0	1.5	0	.62	.03	.26	.6
101	Sebree	86.0	10.0	4.0	79.5	0	1	0	0.5	18.5	0	0.5	0	1.0	0	.30	.02	.12	.6
98	Curlew	90.0	4.5	5.5	70.0	1	0	1	1.0	28.0	0	0	0	1.0	0	.32	.03	.14	.6
99	Above Mining City Coal	87.0	10.0	3.0	94.0	0	1	0	0.5	1.5	0	3.0	0	1.0	0	.46	.02	.15	.6
103	Aberdeen	79.0	9.0	12.0	82.5	1	1	0	1.0	15.5	0	0.5	0	0.5	0	.56	.03	.20	.6
104	Aberdeen	90.0	6.0	4.0	89.0	1	0	0	0.5	10.0	0	0.5	0	0	0	.40	.02	.14	.6
97	Finnie	82.5	11.0	6.5	70.0	2	2	0	2.0	26.0	0	0.5	0.5	0	1.0	.52	.02	.20	.6
95	Grindstaff	96.0	2.5	1.5	86.0	0	0	0	0	13.5	0		0.5	0	0	.24	.03	.10	.6
—*	Average of 17 Caseyville samples	78.8	13.7	7.5	91.2	—	—	—	0.2	7.6	0	0.8	0.2	—	—	—	—	—	
								INDIANA											
278	Merom	72.0	11.5	16.5	74.5	9	3	1	6.5	13.5	0	2.0	3.5	0	0	.70	.03	.30	.5
414	Above Parker Coal	65.0	9.0	26.0	67.0	8	3	1	6.0	25.5	0	0.5	0.5	0.5	0	1.00	.04	.45	.5
413	Inglefield	83.0	7.0	10.0	68.0	6	3	2	5.5	20.0	0	1.0	3.0	1.5	1.0	.36	.02	.14	.6
405	Above No. V Coal	70.0	10.0	20.0	59.0	5	5	2	6.0	23.5	1.5	0.5	3.5	4.5	1.5	.48	.02	.14	.5
415	Petersburg Fm.	77.5	12.5	10.0	69.5	1	1	2	2.0	25.0	0	1.0	0.5	2.0	0	.30	.02	.12	.6
279	Above No. IV Coal	84.0	7.5	8.5	63.5	5	6	0	5.5	25.5	0	0	3.5	2.0	0	.44	.02	.16	—
406	Above No. IV Coal	75.5	12.0	12.5	59.5	7	7	1	7.5	25.5	0	1.0	2.5	3.5	0.5	.36	.02	.14	—
411	"Isabel"	65.0	13.0	22.0	66.0	6	2	2	5.0	21.5	0	1.0	4.5	2.0	0	.66	.04	.36	.5
416	"Isabel"	71.5	7.5	21.0	74.5	5	2	2	4.5	16.5	0	1.5	3.0	0	0	1.16	.06	.54	.6
410	Staunton Fm.	84.0	8.0	8.0	62.0	7	2	2	5.5	24.5	0	1.5	2.0	3.5	1.0	.40	.02	.10	.5
404	Staunton Fm.	80.0	9.0	11.0	71.0	3	1	2	3.0	23.0	0	0	1.0	1.0	1.0	.30	.02	.12	.5
89	Above Minshall Coal	72.5	12.5	15.0	60.0	5	5	1	5.5	29.0	0	1.0	3.5	0.5	0.5	.86	.04	.35	.6
407	"Seville"	76.5	11.0	12.5	69.0	6	5	2	6.5	20.0	0	0	2.0	1.5	1.0	.60	.03	.16	.6
409	Over L. Block Coal	87.5	5.5	7.0	86.0	1	2	1	2.0	9.5	0	0.5	2.0	0	0	.40	.04	.15	.6
412	Over Cannelton Coal	85.0	7.5	7.5	89.5	1	2	0	1.5	7.5	0	0.5	0.5	0.5	0	.48	.03	.20	.6
—*	Average of 18 Mansfield samples	75.3	16.6	8.1	87.3	—	—	—	2.3	8.2	1.0	0.7	0.5	—	—	—	—	—	—

									MISSOURI										
284	Knobtown	90.5	4.5	5.0	50.0	8	4	1	6.5	36.0	3.0	1.0	2.5	1.0	0	.20	.02	.08	—
289	Knobtown	92.0	5.0	3.0	34.0	5	2	0	3.5	55.5	4.5	0.5	1.0	1.0	0	.20	.01	.06	—
282	Hepler	91.5	4.5	4.0	42.5	3	4	0	3.5	45.0	2.0	2.0	1.5	2.0	1.5	.20	.01	.06	—
419	Pleasanton Group	91.0	3.0	6.0	25.0	0	0	0	0	0	75.0	0	0	0	0	5.00	.05	.15	.7
417	Warrensburg	65.5	9.5	25.0	74.5	7	4	2	6.5	6.5	0	5.5	7.0	0	0	1.10	.06	.50	.6
418	Warrensburg	79.5	5.5	15.0	58.0	4	3	2	4.5	19.5	10.5	2.0	4.5	1.0	0	.66	.02	.24	.5
283	Flint Hills	—	—	—	44.0	0	1	0	0.5	47.5	0	0.5	0	5.0	2.5	.07	.01	.03	—
287	Lagonda	86.0	7.5	6.5	58.0	5	2	1	4.0	34.5	0	0.5	0.5	1.0	1.5	.34	.01	.09	—
280	Lagonda	83.0	8.0	9.0	37.5	4	0	0	2.0	41.0	16.0	0	1.0	0.5	2.0	.20	.02	.08	.5
285	Below Bevier Coal	93.5	2.5	4.0	84.0	2	0	0	1.0	12.5	0	0.5	0.5	1.0	0.5	.24	.02	.06	—
398	Below Fleming Coal	85.0	6.5	8.5	63.0	4	1	0	2.5	21.5	7.0	0.5	2.5	2.0	1.0	.22	.01	.08	—
114	Bluejacket	90.5	4.0	5.5	57.5	2	0	0	1.0	36.5	0	1.0	4.0	—	—	.32	.02	.10	.5
115	Bluejacket	88.0	5.0	7.0	93.5	2	4	1	3.5	2.0	0	0.5	0.5	—	—	.52	.04	.22	.5
116	Bluejacket	88.0	5.0	7.0	81.5	2	1	0	1.5	16.0	0	0	1.0	—	—	.90	.02	.30	.6
399	Dry Wood Fm.	87.0	6.0	7.0	62.5	4	1	0	2.5	28.5	0	0.5	3.0	1.5	1.5	.26	.02	.09	.4
400	Warner	86.0	6.5	7.5	80.5	1	0	0	0.5	15.0	0	0.5	3.0	0.5	0	.32	.03	.14	.6
401	Warner	90.5	6.5	3.0	80.5	3	0	0	1.5	14.0	0	0	2.5	1.0	0.5	.22	.03	.10	.6
									APPALACHIAN AND MICHIGAN BASINS										
270	Upper Benwood	70.5	14.5	15.0	64.5	5	3	0	4.0	4.0	23.0	2.5	2.0	0	0	.60	.04	.24	.5
272	Above "bone beds"	80.5	8.0	11.5	51.5	9	2	0	5.5	32.5	1.5	0	1.0	4.0	4.0	.26	.02	.12	—
273	Above Washington "A" Coal	76.5	12.0	11.5	31.5	5	1	1	3.5	19.5	36.0	0	1.5	3.5	4.5	.40	.03	.18	.4
274	Lower Uniontown	77.0	12.5	10.5	40.5	10	5	0	7.5	28.0	20.0	0	2.5	0.5	1.0	.40	.03	.14	.4
275	Lower Sewickley	70.0	15.0	15.0	64.0	7	3	0	5.0	23.5	0	0	1.5	4.0	2.0	.30	.03	.12	—
276	Over Washington Rider "A" Coal	77.5	10.5	12.0	61.5	6	2	0	4.0	25.0	3.5	0.5	2.0	3.0	0.5	.36	.02	.12	.4
271	Ninevah	80.5	10.5	9.0	38.5	4	3	0	3.5	46.5	3.0	0.5	2.0	1.0	5.0	.20	.01	.06	.4
217	Mahoning	62.5	14.5	23.0	95.5	1	2	0	1.5	1.0	0	0.5	1.5	0	0	1.60	.10	.60	.5
220	Above Upper Kittanning Coal	61.5	18.0	20.5	67.5	2	2	0	2.0	26.0	0	1.5	2.0	1.0	0	.56	.03	.24	.5
288	Buck Mountain	44.5	29.0	26.5	68.0	1	0	0	0.5	22.5	0	2.0	6.0	1.0	0	10.00	.06	.64	.5
219	"Leatherwood"	71.0	11.5	17.5	79.0	1	2	5	4.0	13.0	0	1.5	2.5	0	0	1.30	.02	.56	.6
78	Whitesburg	64.0	21.0	15.0	61.5	5	1	1	3.5	29.0	0	0.5	0.5	5.0	0	.64	.03	.26	.4
107	Amburgy	70.5	18.5	11.0	61.5	6	1	6	6.5	31.0	0	0.5	0.5	—	—	.60	.04	.20	.6
218	Above High Splint Coal	59.5	15.5	25.0	80.0	0	0	0	0	13.5	0	3.0	3.5	0	0	2.70	.04	.60	.6
80	"Pottsville"	—	—	—	95.5	—	—	—	0.5	3.0	0	1.0	0	—	—	14.00	.06	—	.8
191	Massillon	74.0	21.5	4.5	91.0	—	—	—	0.5	8.5	0	0	0	—	—	.80	.04	.35	.4
193	Massillon	69.0	16.5	14.5	67.0	—	—	—	5.5	24.0	0	1.0	2.5	—	—	.36	.03	.16	.5
194	Massillon	75.0	14.5	10.5	79.0	4	2	3	4.5	15.0	0	0	1.5	—	—	.60	.03	.22	.6
195	Massillon	50.0	27.0	23.0	95.0	—	—	—	1.0	4.0	0	0	0	—	—	1.20	.08	.58	.7
201	Massillon	62.0	20.0	18.0	82.0	—	—	—	4.0	14.0	0	0	0	—	—	.60	.04	.28	.7
188	Saginaw	78.0	16.0	6.0	95.5	0	1	0	0.5	3.5	0	0	0	0.5	0	.58	.04	.22	.7
79	Bee Rock	64.0	24.5	11.5	97.0	0	3	0	1.5	1.0	0	0.5	0	0	0	6.30	.06	.60	.8
—*	Average of 8 Sharon samples	74.8	14.8	10.4	90.0	—	—	—	1.9	7.0	0	0.6	0.5	—	—	—	—	—	—
—*	Average of 14 Lee samples	75.6	18.8	5.6	85.3	—	—	—	1.4	10.4	2.3	0.5	0.1	—	—	—	—	—	—
—*	Average of 5 Parma samples	73.5	19.7	6.8	86.7	—	—	—	2.3	3.7	4.9	1.0	1.4	—	—	—	—	—	—
222	New River Group	57.0	11.5	31.5	86.5	0	1	0	0.5	10.0	0	1.5	1.5	0	0	1.28	.06	.64	.7
223	Pocohontas Group	66.5	11.0	22.5	79.5	0	0	0	0	17.5	0	1.5	1.0	0	0.5	1.46	.05	.60	.7
215	Pocohontas Group	68.5	12.5	19.0	91.0	0	1	0	0.5	3.5	0	2.0	3.0	0	0	1.26	.06	.45	.7

* For individual analyses, see Siever and Potter (1956).
† The modal analyses are from Potter and Glass (1957); quartz variety and size analyses my own.
* For individual analyses, see Siever and Potter (1956).
* For individual analyses, see Siever & Potter (1956).
* See Siever and Potter (1956).

TABLE 1—(*continued*)
Averages of Post-Caseyville Sandstones by Province

Province	No. of samples	Quartz			Feldspar				Clay	Carbonate	Chert	Rock fragments	Micaceous minerals		Size		
		Igneous	Metamorphic	Meta-quartzite	Total	Plagioclase	Orthoclase	Microcline	Total				Muscovite	Biotite-Chlorite	Maximum	Minimum	Approx. mode	Average roundness
Western Kentucky	19	87.0	6.9	6.1	74.2				2.2	19.8	1.6	0.6	0.2	1.2	0.2			.18
Indiana	15	76.6	9.6	13.8	69.3				4.8	20.7	0.1	0.8	2.4	1.5	0.4			.23
Western Illinois	22	88.3	7.7	4.0	65.1				2.6	25.3	3.6	0.8	0.6	1.3	0.8			.12
Northern Illinois	4	74.0	10.5	15.5	58.8				5.0	27.3	6.8	0.2	0.6	1.1	0.2			.16
Eastern Illinois	6	75.2	11.7	13.1	49.9				5.2	28.8	9.3	0.6	1.6	2.2	2.2			.14
Southwestern Illinois	11	77.6	9.0	13.4	66.8				3.7	17.4	6.2	0.3	1.1	2.8	0.8			.17
Southern Illinois	22	76.2	10.8	13.0	62.9				2.4	18.2	10.2	0.2	0.3	1.6	0.4			.10
New Haven core	25	86.2	8.7	5.1	76.4				2.6	13.8	5.4	0.6	0.1	0.6	0.6			.18
Edwards Co. cores	8	88.1	7.9	4.0	71.3				3.0	17.0	3.9	0.6	1.5	1.6	1.1			.20
Appalachian	22	68.5	16.7	14.8	71.3				3.0	17.0	3.9	0.6	1.5	1.6	1.1			.31
Missouri	17	86.7	5.6	7.7	60.4				2.6	25.4	6.9	0.9	2.1	1.0	0.7			.14

types can be drawn more definitely in the foreseeable future.

In many parts of the northern mid-continent there does not seem to be any definite orthoquartzite basal zone (with the exception of the very local Graydon beds of Missouri, primarily a reworked rubble on the pre-Pennsylvanian unconformity and some basal sandstones of southeastern Iowa[5]). All of the lower sandstones of this province have very nearly the same character—corresponding in type to the transition zone of the Eastern Interior Coal Basin. The basal orthoquartzitic zone is well recognized in the Appalachian and Michigan Basins, however, (Sharon, Lee, Bee Rock, Pocahontas Sandstones). In the eastern provinces the situation with respect to the orthoquartzites seems to be quite similar to that in the Eastern Interior Coal Basin; the zone appears to be somewhat transgressive from south to north, and the upper boundary, although not sharp in all places, generally can be distinguished.

The boundary between the transition and subgraywacke zones in Missouri is approximately at the level of the Lagonda Sandstone, slightly higher than in western Illinois. There are too few analyses to indicate this boundary definitely in the Appalachian Basin but there is a suggestion that the boundary is approximately at the same stratigraphic level as in southern Illinois and western Kentucky.

Within each sandstone type there tend to be certain systematic variations. The younger sandstones of the transition group approach subgraywackes in composition and the older are more similar to the basal orthoquartzites. The younger sandstones of the subgraywacke group tend to have higher proportions of feldspar and mica, although the few analyses of sandstones highest in the section (above the Mt. Carmel Sandstone) suggest that feldspar decreases again.

The averages of modal analyses of all of the post-Caseyville sandstones of the several geographic provinces (table 1) show that there are no pronounced differences among any of the provinces. The same homogeneity is present in the averages of quartz varieties. This widespread homogeneity of the major constituents of the post-Caseyville sand-

[5] H. R. Wanless, personal communication.

stones, extending from, at least, eastern Kansas and southwestern Missouri to northeastern parts of the Appalachian Basin, can be contrasted with the basal orthoquartzite zone, which shows significant differences between western Illinois and other parts of the Eastern Interior Coal Basin (Siever and Potter, 1956, p. 322).

This homogeneity is not complete, however. Western Illinois and Missouri have lower proportions of metamorphic quartz varieties than any of the other provinces, and the Appalachian Basin has higher amounts of metamorphic quartz varieties than any of the basins to the west. Several of the Appalachian sandstones contain more than 40 percent metamorphic quartz varieties.[6] Indiana, northern Illinois, and eastern Illinois have slightly higher amounts of feldspar than other areas but this may be explained in part by the fact that the samples from those provinces are weighted towards the upper part of the section, which everywhere tends to contain greater amounts of feldspar.

Qualitative observations on the characteristics of major and minor constituents of the sandstones reveal further inhomogeneities as well as information bearing on the composition of source rocks.

Although only three varieties of quartz were counted, other types of detrital quartz were noted, in particular, fragments of sedimentary quartzite and second-cycle grains (grains that show abraded secondary overgrowths indicative of a previous sedimentary history). Although the two varieties of sedimentary derived quartz are rarely abundant, some few grains are present in almost all thin sections. Because original detrital outlines of many quartz grains are not recognizable it is probable that there are actually more second-cycle grains present than were noted. Sedimentary quartzite grains are more abundant in the Pennsylvanian sandstones of the Appalachian Basin than in any of the other provinces.

The relative abundance of the three varieties of feldspar counted do not show any great variation either within or between provinces, although the number of grains actually counted was too small to warrant very firm conclusions. Plagioclase is more abundant than orthoclase, and microcline is far less abundant than either. In only one province, Indiana, was microcline present consistently throughout the profile. The generally small proportions of microcline indicate either a characteristic of particular Pennsylvanian source areas or perhaps a general order of abundance of feldspar characteristic of many sediments. Both fresh and weathered feldspars are present in most of the sandstones. Weathered grains show extensive kaolinization and in some cases alteration to a micaceous type of clay mineral. Although both kinds of grains are commonly present there is a slight tendency for a greater amount of fresh feldspar high in the section. There seems to be no obvious systematic regional variation in the proportion of weathered grains.

Clay minerals in the Pennsylvanian sandstones have not been analyzed in detail in this investigation. Several clay analyses by H. D. Glass, as well as previous studies (Glass, Potter, and Siever, 1956; Potter and Glass, 1957) indicate that kaolinite, mica (illite), chloritic clay minerals, and mixed-layer lattice minerals are commonly present. Kaolinite and illite are most abundant in the Caseyville orthoquartzite group; chloritic clay increases in the transition group and reaches a maximum in the subgraywacke group. In the upper subgraywacke group a biotitic clay mineral is present, sometimes in fairly large amounts. In several thin sections a brownish, pleochroic, micaceous clay mineral (biotite) grades into a greenish, very low birefringent clay mineral (chlorite).

The clay minerals, identified originally from X-ray diffraction patterns, were tentatively identified in thin section. Kaolinite can be distinguished easily by its index of refraction, low birefringence, and characteristic vermicular habit (seen only when crystals are of sufficient size). Micaceous and mixed-layer clays can be recognized by medium to moderately high birefringence and eminent cleavage. Chloritic clay is characteristically very slightly birefringent and colored various shades of green, often pleochroic. A grass-green and yellowish green, microcrystalline, low birefringent clay was

[6] In part this effect may be due to increased amounts of strained quartz as a result of post-Pennsylvanian deformation.

TABLE 2.—*Location of samples*

No.	Location	No.	Location	No.	Location
9	Jasper Co., Ill.	247	Knox Co., Ill.	327	Clark Co., Ill.
57	Williamson Co., Ill.	248	Knox Co., Ill.	328	Saline Co., Ill.
58	Saline Co., Ill.	249	Bureau Co., Ill.	329	Williamson Co., Ill.
59	Saline Co., Ill.	250	Peoria Co., Ill.	330	Franklin Co., Ill.
60	Williamson Co., Ill.	251	Fulton Co., Ill.	353	Clay Co., Ill.
61	Saline Co., Ill.	252	Fulton Co., Ill.	398	Vernon Co., Mo.
62	Saline Co., Ill.	253	Knox Co., Ill.	399	Vernon Co., Mo.
63	Saline Co., Ill.	254	Macoupin Co., Ill.	400	Jasper Co., Mo.
64	Saline Co., Ill.	255	Macoupin Co., Ill.	401	Vernon Co., Mo.
65	Pope Co., Ill.	256	Macoupin Co., Ill.	404	Spencer Co., Ind.
74	Jackson Co., Ill.	257	Macoupin Co., Ill.	405	Daviess Co., Ind.
75	Jackson Co., Ill.	258	Effingham Co., Ill.	406	Green Co., Ind.
76	Jackson Co., Ill.	259	Vermilion Co., Ill.	407	Spencer Co., Ind.
78	Perry Co., Ky.	260	Vermilion Co., Ill.	409	DuBois Co., Ind.
79	Wise Co., Va.	261	Saline Co., Ill.	410	Spencer Co., Ind.
80	Northumberland Co., Pa.	262	Muhlenberg Co., Ky.	411	Daviess Co., Ind.
82	Fulton Co., Ill.	263	Webster Co., Ky.	412	Perry Co., Ind.
83	Fulton Co., Ill.	264	Webster Co., Ky.	413	Gibson Co., Ind.
84	Fulton Co., Ill.	266	Hopkins Co., Ky.	414	Knox Co., Ind.
85	Fulton Co., Ill.	267	Webster Co., Ky.	415	Sullivan Co., Ind.
89	Parke Co., Ind.	268	Webster Co., Ky.	416	Spencer Co., Ind.
95	Union Co., Ky.	269	Hopkins Co., Ky.	417	Marion Co., Mo.
97	Union Co., Ky.	270	Washington Co., Pa.	418	Carroll Co., Mo.
98	Butler Co., Ky.	271	Wetzel Co., W. Va.	419	Sullivan Co., Mo.
99	Butler Co., Ky.	272	Washington Co., Pa.	C-1319B	Edwards Co., Ill., depth 2080′
101	Webster Co., Ky.	273	Washington Co., Ohio	C-1346A	Edwards Co., Ill., depth 1656′
103	Butler Co., Ky.	274	Belmont Co., Ohio	C-1346B	Edwards Co., Ill., depth 1482′
104	Butler Co., Ky.	275	Washington Co., Pa.	C-1360A	Edwards Co., Ill., depth 1356′
107	Leslie Co., Ky.	276	Washington Co., Ohio	C-1360B	Edwards Co., Ill., depth 1207′
108	Rock Island Co., Ill.	278	Sullivan Co., Ind.	C-1434	Jackson Co., Ill., depth 1283′
109	Rock Island Co., Ill.	279	Sullivan Co., Ind.	C-1465A	Edwards Co., Ill., depth 1843′
114	Cherokee Co., Kansas	280	Boone Co., Mo.	C-1465B	Edwards Co., Ill., depth 1761′
115	Vernon Co., Mo.	282	Livingston Co., Mo.	C-1466	Edwards Co., Ill., depth 331′
116	Henry Co., Mo.	283	Randolph Co., Mo.	4-12	New Haven core, White Co., Ill., depth 355′
157	McDonough Co., Ill.	284	Ray Co., Mo.	4-16	New Haven core, White Co., Ill., depth 402′
158	Henderson Co., Ill.	285	Randolph Co., Mo.	4-50	New Haven core, White Co., Ill., depth 601′
163	Warren Co., Ill.	287	Boone Co., Mo.	4-60	New Haven core, White Co., Ill., depth 657′
164	Knox Co., Ill.	288	Northumberland Co., Pa.	4-60C	New Haven core, White Co., Ill., depth 676′
188	Shiawassee Co., Mich.	289	Livingston Co., Mo.	4-70	New Haven core, White Co., Ill., depth 725′
191	Perry Co., Ohio	291	Williamson Co., Ill.	4-71	New Haven core, White Co., Ill., depth 735′
193	Perry Co., Ohio	292	Saline Co., Ill.	4-87	New Haven core, White Co., Ill., depth 816′
194	Geauga Co., Ohio	297	Edgar Co., Ill.	4-97	New Haven core, White Co., Ill., depth 851′
195	Portage Co., Ohio	301	Saline Co., Ill.	4-110	New Haven core, White Co., Ill., depth 902′
201	Vinton Co., Ohio	303	Williamson Co., Ill.	4-119	New Haven core, White Co., Ill., depth 943′
206	St. Clair Co., Ill.	304	LaSalle Co., Ill.	4-127	New Haven core, White Co., Ill., depth 974′
215	Fayette Co., W. Va.	305	Effingham Co., Ill.	4-140	New Haven core, White Co., Ill., depth 1005′
217	Grant Co., W. Va.	306	Tazewell Co., Ill.	4-151	New Haven core, White Co., Ill., depth 1076′
218	Letcher Co., Ky.	307	Jefferson Co., Ill.	4-155	New Haven core, White Co., Ill., depth 1100′
219	Letcher Co., Ky.	308	LaSalle Co., Ill.	4-160	New Haven core, White Co., Ill., depth 1128′
220	Somerset Co., Pa.	309	Clark Co., Ill.	4-168	New Haven core, White Co., Ill., depth 1141′
222	Fayette Co., W. Va.	310	Jackson Co., Ill.	4-182	New Haven core, White Co., Ill., depth 1248′
223	Fayette Co., W. Va.	319	Jackson Co., Ill.	4-183	New Haven core, White Co., Ill., depth 1251′
230	Union Co., Ky.	320	Tazewell Co., Ill.	4-190	New Haven core, White Co., Ill., depth 1312′
231	Union Co., Ky.	321	Peoria Co., Ill.	4-202	New Haven core, White Co., Ill., depth 1365′
233	Webster Co., Ky.	322	LaSalle Co., Ill.	4-213	New Haven core, White Co., Ill., depth 1393′
234	Webster Co., Ky.	323	Edgar Co., Ill.	4-219	New Haven core, White Co., Ill., depth 1437′
235	Webster Co., Ky.	324	Christian Co., Ill.	4-240A	New Haven core, White Co., Ill., depth 1630′
245	Knox Co., Ill.	325	Livingston Co., Ill.	4-251	New Haven core, White Co., Ill., depth 1757′
246	Fulton Co., Ill.	326	Saline Co., Ill.		

tentatively identified as glauconitic clay. A rough estimate, from thin sections, of the relative abundance of these clay types in the various provinces suggests that Missouri and northern, western, and southwestern Illinois contain more biotitic and chloritic clay than do Indiana, southern Illinois, western Kentucky, and the Appalachian Basin.

Large amounts of kaolinite in many of the sandstones occur as "books" or vermicular aggregates and appear to be authigenic; this feature of the kaolinite also has been noted in the Caseyville Group (Glass, Potter, and Siever, 1956) and in the underlying sandstones of the Upper Chester Series in Illinois (Siever, 1953). Other than the gradation between biotitic and chloritic clay it was not possible from microscopic study to evaluate definitely any authigenesis of other types of clay minerals. It does appear, however, in a number of thin sections that a micaceous clay mineral has replaced marginal areas of detrital quartz grains.

Carbonate minerals are common in the sandstone, primarily as cement, but in some

few sandstones as detrital fossil fragments. Calcite is by far the most abundant with smaller amounts of dolomite, ferroan-dolomite, and siderite. Some thin sections show "floating" quartz grains completely surrounded by large masses of calcite; undoubtedly much of this appearance is caused by extensive post-depositional replacement of quartz grains, but some few appear to be the result of recrystallization of calcite that originally was in the form of detrital grains. The effects of diagenesis and cementation of carbonate minerals in these sandstones is not discussed here but will be the subject of another study.

Chert fragments are present in most of the sandstones, normally constituting no more than 1% of the rock but in some samples rising to 4 and in one case 5.5%. Chert is, of course, just a particular variety of rock fragment, and in most sandstones where the chert is more abundant, so also are other rock fragments. Most of the chert is nonfossiliferous but occasional grains show some fossils. There are no distinctive features of the chert that would make it possible to guess the identity of the source rocks other than the assumption that the erosion of a limestone terrain was responsible.

Rock fragments are found in approximately the same amounts and distribution as chert. Almost all rock fragments are of unmetamorphosed argillaceous and silty argillaceous sediments. The fragments are similar to Pennsylvanian siltstones and shales. Some show slight foliation indicative of mild metamorphism. A quite different type of sedimentary and metamorphic rock fragment is found in some samples in the Appalachian Basin (in particular those from the Pochontas Group), in several samples of the Isabel sandstone in the Eastern Interior Coal Basin, and in samples of the Warrensburg Sandstone of Missouri. These rock fragments are of well indurated argillaceous and silty rocks that obviously were derived from a metasedimentary terrain. In the Pocahontas samples they are associated with fragments of sedimentary and metamorphic quartzite. In the Isabel and Warrensburg Sandstones they are associated with a larger number of chert fragments. Conglomerates at the bottom of channel sandstones commonly contain many fragments of clay ironstones, siderite concretions, and coal; these fragments are almost certainly of Pennsylvanian age and were locally derived.

Flakes of micaceous minerals are present in many of the sandstones of the transition group and almost all the sandstones of the subgraywacke group. Muscovite is almost always more abundant than biotite and chlorite combined. Because only the large, definitely detrital mica flakes were counted, in order to avoid confusion with the micaceous minerals of the clay matrix, the relative abundances of micas probably are underestimated. Some geographic differentiation is possible on the basis of the ratio of biotite to chlorite. The Appalachian Basin, Indiana, and eastern and northern Illinois have much more biotite than chlorite; western Kentucky and southwestern and southern Illinois have slightly more chlorite than biotite; western Illinois and Missouri have equal amounts. Many flakes of biotite show various kinds of alteration—slight loss of pleochroism, partial decolorization, and alteration to chlorite. In a number of flakes, some of the laminae are altered and others are not. The nonuniform nature of the alterations suggests that they date from weathering processes in the source area rather than being penecontemporaneous or diagenetic. The presence of these alterations in unweathered core samples shows that post-Pennsylvanian weathering was not responsible.

Iron oxide minerals, abundant in many of the outcrop samples, were ignored in the analyses.

Several other minerals are present in minor amounts. Barite is found as cement in scattered samples in several provinces. In the few thin sections where it constituted a significant amount, it was included with carbonate minerals. Glauconite grains were found in many of the subgraywackes and in some of the transition group sandstones. The grains range from grass-green to yellowish-green. Glauconite may be common, as in the McLeansboro Group sandstones of northern and western Illinois, but is never abundant. It is difficult to tell whether these grains were derived from older glauconitic sediments or originated in situ in the Pennsylvanian sediments. Militating against the former explanation is the fact that glauco-

TABLE 3.—*Heavy minerals*
F = Flood (>30%); A = Abundant (>10%); P = Present; X = not present

Sample	Zircon	Tourmaline	Rutile	Garnet	Chlorite
SOUTHERN ILLINOIS					
Over Greenup ls., Cumberland Co.	F	A	P	F	X
Mt. Carmel ss., White Co.	F	A	P	P	X
Shoal Creek ss., Washington Co.	A	F	P	P	X
Shoal Creek ss., Saline Co.	A	A	P	A	A
Under Scottville ls., Macoupin Co.	A	A	X	A	P
Anvil Rock ss., Gallatin Co.	F	A	P	X	P
Over No. 4 coal, Jackson Co.	F	A	P	P	P
Palzo ss., Saline Co.	F	A	P	X	X
Palzo ss., Edwards Co.	F	A	P	P	P
Delong ss., Macoupin Co.	F	A	P	P	A
Curlew ss., Saline Co.	F	A	P	X	X
Delwood ss., Jackson Co.	F	A	P	X	X
WESTERN AND NORTHERN ILLINOIS					
Gimlet ss., Peoria Co.	A	A	P	A	A
Gimlet ss., Livingston Co.	A	F	P	A	P
Copperas Creek ss., LaSalle Co.	F	A	P	A	A
Vermilionville ss., LaSalle Co.	A	A	P	A	P
Pleasantview ss., Greene Co.	F	A	X	A	P
Isabel ss., Fulton Co.	F	A	P	A	P
Seville ss., Mercer Co.	F	A	A	P	X
Pope Creek ss., Rock Island Co.	F	A	A	P	X
Tarter ss., Rock Island Co.	F	A	P	X	X
EASTERN ILLINOIS AND INDIANA					
Merom ss., Sullivan Co., Ind.	A	F	X	A	X
Below Livingston ls., Crawford Co., Ill.	A	F	P	A	A
Above "Hunter" coal, Edgar Co., Ill.	A	A	P	P	P
30′ below Livingston ls., Vermilion Co., Ill.	A	F	P	A	A
Above Lonsdale ls., Vermilion Co., Ill.	A	A	P	F	X
Above No. IV coal, Parke Co., Ind.	A	A	X	P	A
Above Minshall coal, Parke Co., Ind.	P	A	P	A	P
WESTERN KENTUCKY					
Dixon ss., Webster Co.	F	A	P	X	X
Above No. 14 coal, Hopkins Co.	F	F	A	X	P
Below No. 11 coal, Webster Co.	F	A	X	P	P
"Cuba" ss., Hopkins Co.	F	A	P	X	X
Sebree ss., Webster Co.	F	F	P	X	P
Curlew ss., Butler Co.	F	A	X	X	P
Lower Aberdeen ss., Butler Co.	A	F	P	X	X

nite is relatively unstable mechanically and would not be expected to survive the rigors of transport.

Faunal remains are uncommon in the sandstones of all basins studied. Fragmentary remains, unidentifiable, were found in samples of the Pope Creek and Babylon of western Illinois, the Vanderburg of western Kentucky, and the Battery Rock of southern Illinois. In addition, identifiable fragments of crinoids and molluscs are abundant in the Boskydell of southern Illinois, and poorly preserved remains of bryozoa were found in the Curlew and Delwood of southern Illinois (identified by C. W. Collinson of the Illinois State Geological Survey). Several other sandstones contained occasional fragments that may be fossil fragments but could not be determined as such.

Plant fossils are common in many of the sandstones. As was noted for the Trivoli Sandstone (Siever, 1949, p. 617) most of them are fragmentary, fusainized or vitrinized shreds and particles, rarely identifiable as to family. Most of them probably are woody remains. In some sandstones in south-

ern Illinois and the Appalachian Basin, layers of humic matter, similar to that found in normal attrital coals, were noted. Lenticular beds of coal, normally lacking an underclay, are found in some channel sandstones.

By far the most abundant heavy minerals (table 3) are zircon and tourmaline. Others, such as rutile, garnet, anatase, apatite, biotite, and chlorite, are common. Most of the heavy minerals do not show any great variation throughout the Pennsylvanian, nor do they vary greatly from one part of the basin to another. The most notable exception is garnet, which Willman (1928) early noted as abundant in western Illinois only in the Carbondale and McLeansboro Groups, but rare—only a few scattered grains being found—in older sandstones. As shown in table 3, garnet abundance varies not only with stratigraphic position but also with geographic position in the Eastern Interior Coal Basin. Only a few grains of garnet were found in any of the sandstones of western Kentucky. Only three sandstones high in the McLeansboro Group in southern Illinois showed garnet abundant. In contrast, all of the Carbondale and McLeansboro sandstones of western and northern Illinois showed abundant garnet and almost all of the samples from eastern Illinois and Indiana, from the Tradewater Group up, show abundant garnet. On the basis of the samples studied it seems that garnet appeared in abundance first in Indiana during the early part of middle Tradewater time, slightly later in western and northern Illinois at the beginning of Carbondale time, and reached southern Illinois in any abundance only late in the period, in McLeansboro time. Chlorite abundance tends to follow the same pattern of stratigraphic position as garnet, but more erratically.

The great majority of the modal sizes of the sandstones, as shown in figure 6, are in the fine and very fine sand class; only two samples have modal sizes in the coarse sand range. Maximum sizes in the nonconglomeratic samples range up to about 1.5 mm but tend to cluster between 0.2 and 0.5 mm. The tabulation of minimum size is not too meaningful, since the lower limit of detrital particle size actually extends down into clay sizes. In general the sands are fairly well sorted if the clay matrix is ignored. Averages

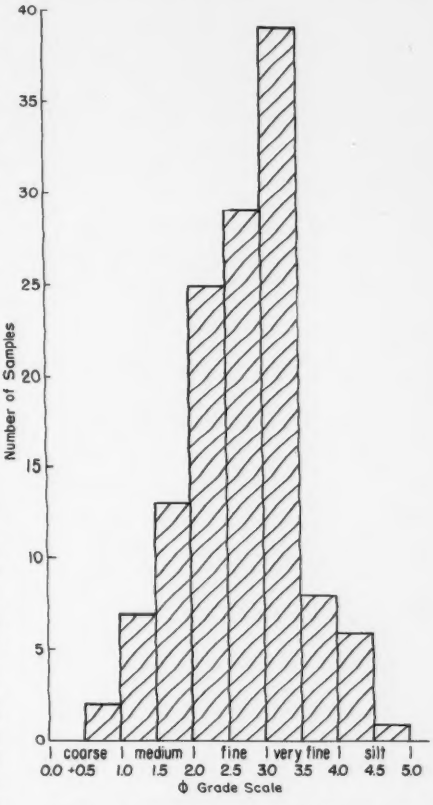

FIG. 6—Histogram of modal sizes of Pennsylvanian sandstones.

of modal sizes by province indicate that the larger sizes are found in the Appalachian Basin, southern Illinois, western Kentucky, and Indiana. The smaller modal size averages are from Missouri, western, northern, and eastern Illinois. Within each province there is a general tendency for the upper sandstones to be finer than the lower ones.

Figures given in table 1 for roundness (of quartz) are not very reliable because of the obscuring of original detrital grain boundaries by secondary enlargement and replacement by cement or matrix. All indications are, however, that a high proportion of the grains are well rounded. Qualitative observation suggests that there is a mixture of two

populations of quartz grains, one group of roundness 0.5 to 0.7 and another group of 0.3 to 0.5. Caution is necessary in the interpretation of the roundness figures given in the analyses, for it is obvious that quartz roundness is very closely related to size. Considering the size factor, it appears that western Illinois and Missouri have rounder grains than the other areas.

SEDIMENTATION

Stratigraphic and petrographic observations give some evidence of the conditions of sedimentation of the sandstones, namely the physical, chemical, and biological factors acting on the transported detritus at the site of deposition. These observations alone, however, give basis for only tentative conclusions as to the environment of sedimentation, which is at present operationally defined to a large degree in terms of the geomorphology of the site of deposition, for example, estuary, tidal lagoon, or valley flat. This is true partially because it is questionable how different the impress of environmental factors of closely related environments, such as those associated with a shoreline, is on the mineralogy and texture of sand. Another reason for the difficulty in reconstructing the geomorphology of the depositional site is that as yet we have insufficient knowledge of three-dimensional stratigraphic patterns associated with modern sites of deposition. Therefore, it is within these limitations that we must interpret Pennsylvanian environments of deposition.

Evidence from grain-size distributions, sedimentary structures, and limited lateral persistence as a lithologic unit of any particular sandstone lens indicates that current action varied greatly in intensity over relatively small areas. Length of time that any particular level of current activity persisted at any one place must have been short, as indicated by the relatively thin units of uniform sandstone lithology of all but some channel sands. It is obvious that current strength and variability were greater in those areas where the coarser, more heterogeneous channel sands were laid down than in the much broader areas that were the sites of sheet-sand deposition. The current variants and channel-sheet contrasts, however, are superimposed on a general homogeneity of texture, mineralogy, and stratigraphic pattern over the entire basin. In general lithologic aspect, the sandstones are homogeneous over several now separated basins from Missouri to Pennsylvania. Thus we have a picture of current activity whose intensity varied greatly in time and space within certain average limits. But the limits and the degree of variability were remarkably uniform over much of the north central and northeastern part of the United States.

To be sure, the limits are broad. Only the extremes of current intensity, and the types of sandstone beds resulting therefrom were lacking, and it is easier to eliminate the missing than to enumerate the present. If we neglect the volumetrically insignificant basal conglomerates of certain channel sandstones, we find that sandstones reflecting extremely vigorous current action are, by and large, not present, as evidenced by the overwhelming predominance of medium, fine, and very fine-grained sandstones. Evidences of turbidity current action are rare and it must be presumed that during the Pennsylvanian period it was exceptional to have submarine slopes steep enough for this mechanism (Potter and Glass, 1957). The rarity of turbidity current structures, the abundance of plant debris in so many sandstones, and the abundant evidence of current activity in deposition and reworking of sediments all suggest that water was not deep. It is doubtful if the evidence at hand will permit any more precise evaluation of depth of water.

The general fineness of grain size in western Illinois and Missouri in comparison with the coarser grain distributions of other areas in the Eastern Interior Coal Basin is fair evidence that average currents were weaker in the western Illinois shelf and northern mid-continent areas than in areas to the east and south. And yet the fact that the amount of interstitial clay is roughly the same in all areas shows that the kinds of transporting currents were similar insofar as their influence on sorting is concerned.

Evidence is inconclusive on the knotty question of the marine or nonmarine origin of the sandstones. In essence this question is not concerned with the difference between saline and fresh waters but with discrimination between terrestrial and submarine or

shore-line environments. A few sandstones contain marine fossils and so are easily placed. A greater number contain stigmarian roots throughout the sandstone and are almost certainly nonmarine. But the great bulk of sandstone beds show few features that would place them incontrovertibly into one category or another. Weller (1930) summarized many of the evidences for nonmarine origin and no really new arguments have been brought out since. Perhaps the strongest evidence is the lack of marine fossils in most of the sandstones. The fact that channel cutting may have been subaerial is actually not an argument for nonmarine origin of the sandstones filling the channels, for it is entirely possible that the fills were deposited by marine inundations subsequent to channel erosion.

Evidence pointing to marine origin is just as weak as that suggesting nonmarine conditions. In spite of extensive field and petrographic observation, very few of the sandstones yield any marine fossils. Although the wide lateral persistence of certain sheet-sand zones may be the product of shallow marine current action, the fact that so many sandstones show abrupt and extreme lateral variation vitiates this argument. A more forceful argument lies in the excellent mixing and homogenization of the detritus. It is probable such extensive mixing of detritus is more likely a result of shallow marine sediment distributive processes than terrestrial (alluvial) distribution. Here again, more definite conclusions must await more intensive mineralogic and petrographic studies of modern analogues.

It is possible to harmonize the somewhat conflicting evidence into a unified hypothesis. This hypothesis is that the sandstones were deposited in a variety of shallow water, littoral, lagoonal, deltaic, and coastal plain alluvial environments in which the time rate of migration of the strand line was high. Thus characterization of the specific environment of these ancient sediments, whose original depositional features are so uninformative in terms of possible analogues, can only be a time average of a dynamic situation involving the rapid migration of the shore line over great distances in short times. Thus, in terms of present knowledge, the characterization of the environment of these sandstones at a point in both space and time becomes almost indeterminate.

SOURCE AREAS

The point of departure for the following consideration of source areas is the paleogeographic synthesis proposed for the basal Pennsylvanian of the north-central and northeastern states by Siever and Potter (1956). According to this synthesis, in earliest Pennsylvanian time the major source areas were parts of the Canadian Shield to the north and a linear uplifted part of the Appalachian mobile belt to the east and northeast, both of which contributed detritus that was transported down the regional slope of the craton from northeast to southwest. A minor source was the transcontinental arch in Minnesota, the Dakotas, and Colorado that contributed material to western Illinois.

In the part of the continent considered here there is no evidence of any major orogeny or epeirogeny in the early Pennsylvanian time that would have led to any drastic changes in relationships of source areas to general directions of sediment flow and basins of deposition. Hence it is likely that the paleogeographic situation in early Pennsylvanian time persisted, with some modifications, throughout the entire Pennsylvanian period.

The differences in composition and texture between the basal Pennsylvanian sandstones of western Illinois (and Missouri) and other areas to the south and east is pronounced. In the sandstones higher in the Pennsylvanian section the difference persists, though it is less obvious. Homogenization and mixing of sediment was extensive in the early Pennsylvanian. It was even more extensive in the later Pennsylvanian.

Limited observations of cross-bedding suggest that general directions of sediment transport to the south and southwest persisted from early Pennsylvanian time throughout the entire period. The greater homogeneity of the later Pennsylvanian sandstones, the lessened differences between provinces, and the similarity of detritus of Eastern Interior and Appalachian Basins implies that eastern and northeastern sources supplied an increased amount of sediment after early Pennsylvanian time,

overwhelming the smaller amounts coming from the Canadian Shield and transcontinental arch. The smaller relative contributions of northern sources may be due not only to dilution by material derived from the east and northeast but also to gradual transgression and overlapping of Pennsylvanian sediments onto the arch and the southern parts of the Canadian Shield.

The changes in composition of the sandstones give a clue to changes in source-area terrains. The succession of orthoquartzite group, transition group, and subgraywacke group can be taken as a reflection of continuing and deeper erosion of the sedimentary cover of the source area. The persistence throughout the entire Pennsylvanian of western Illinois of a small amount of well-rounded tourmalines shows that there was a continuing small source of sedimentary rocks with a long abrasion history, including several sedimentary cycles, presumably the same source rocks that contributed very mature materials to the basal Pennsylvanian of the same area. The presence of second-cycle quartz grains in all areas indicates that earlier sediments were a continuing source for the Pennsylvanian sands of all of the basins. But the decreased quantity of the multicycle earlier sediments in the upper Pennsylvanian of all areas reflects the lessened importance of these sediments as source rocks. Their decreased importance as source rocks probably is due to a combination of overlapping by Pennsylvanian sediments, attrition by continued erosion in the source area, and dilution in the basin of deposition by materials from other sources.

The higher content of less stable minerals —feldspar, biotite and chlorite—in middle and upper Pennsylvanian sandstones marks the partial unroofing of some source areas that exposed metamorphic and/or igneous terrains. The common association of weathered and fresh feldspars in the sandstones of all areas would seem to point toward a source terrain in which weathered metamorphic and/or igneous rocks gradually were being cut through to expose fresh rocks of the same type. The generally high ratio of weathered to unweathered feldspar probably reflects a deeply weathered zone that was completely eroded only in small areas.

Garnet, abundant in the middle and upper Pennsylvanian of western Illinois but rare to the south and east, must have come from metamorphic or igneous rocks to the north and probably northwest of western Illinois that became exposed in significant quantity only towards the end of lower Pennsylvanian time. The absence of garnet in western Kentucky reinforces the conclusion that detritus from the northwest that was deposited in western Illinois was not transported very far south, or if so, was vastly diluted by non-garnet bearing detritus from the east and northeast. The systematic mapping of the garnet zone in all the sedimentary basins is a future study that should be rewarding in detailing the patterns of contributions of various sources to the several basins.

The composition of even the uppermost Pennsylvanian sandstones of the basins studied does not indicate that the source areas were completely stripped of their sedimentary cover. The youngest sandstones studied, those of the Dunkard Series in West Virginia, Ohio, and Pennsylvania (Cross, 1950), are similar in every respect to the older subgraywacke sandstones and do not represent any significantly increased tectonism. Nowhere in the areas studied are there any sediments, such as arkoses or true graywackes (high-rank graywackes of Krynine, 1948), that could be interpreted as the products of intense tectonic activity. It must be concluded that any such tectonic sediments that may have been associated with the later stages of Appalachian orogeny were produced later than the early Permian and have since been eroded or obliterated.

The petrographic composition of certain channel sandstones, notably the Isabel of western Illinois and the Warrensburg of Missouri, which is distinctly different from that of the average sandstone in these districts, implies that during some intervals of channel cutting, the normal pattern of sediment transport was interrupted. It is possible but not probable that the detritus in the sandstones came from different and somewhat unique sources. It is more likely that the detritus of these sands represents material from sources due north and northwest that, due to temporary alterations in regional slope directions, was undiluted by material from other sources.

REGIONAL TECTONICS

The major conclusion as to tectonism in source areas, from a consideration of the composition of the detritus, is that the sources were subjected to epeirogenic uplifts and some local mild orogeny, but at no time in the Pennsylvanian Period were they involved in major orogenic episodes. As in early Pennsylvanian time, major tectonic activity was restricted to areas in the southwest and far west that had little direct influence on sedimentation in the basins here considered.

Tectonic activity on the craton was limited to continuing intermittent subsidence of basins such as the Eastern Interior, separated from adjoining basins by quiescent arches or low swells that were no barrier to sediment flow but simply subsided less rapidly and so received less sediment. Within the basins were slowly growing structural elements, synclines and anticlines, inherited from upper Mississippian times, that influenced mainly the thickness rather than the kind of deposition. The fact that there was little relation other than thickness between inter- and intra-basin structures and sedimentary facies indicates that sedimentation kept pace with subsidence and that there was little topographic expression of the structural elements.

Thus the general aspect of the north-central and northeastern states, and source areas to the north, northeast, and east, during the entire Pennsylvanian Period, was one of only moderate tectonic activity. The history of source area movements can be traced by the composition of the sediments in the Eastern Interior Coal Basin. Starting with the upper Mississsippian Chester Series, for the first time since the early Paleozoic and as a result of source area uplifts, large volumes of coarse clastics were deposited in the basin. During Chester time mainly orthoquartzitic sands were laid down (Siever, 1953) and, after the interruption of the Mississippian-Pennsylvanian erosion interval, deposition of clean quartz sands continued into early Pennsylvanian time. Shortly thereafter, continued tectonic activity in the source areas resulted in more extensive erosion of sedimentary rocks, which themselves were one or two cycles removed from ultimate igneous and metamorphic derivation. This activity combined with more active downwarping of the sedimentary basins to produce the transition-group sands and later the subgraywackes. As erosion continued in middle and upper Pennsylvanian time more of the sedimentary cover of the sources was removed and small areas of metamorphic and igneous terrain were exposed. Here the sedimentary record stops and one can only speculate on the probable increasing tectonic character of the sediments associated with the early orogenic phases of the final stage of the Appalachian revolution, the mountain building that destroyed the geosyncline and welded it to the stable core of the continent.

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FIGS. 1–9, SEPTEMBER, 1957

STRUCTURAL PETROLOGY OF THE PICO ANTICLINE, LOS ANGELES COUNTY, CALIFORNIA[1]

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ABSTRACT

A petrofabric field and laboratory study was made of the rocks of the Pico Anticline, Los Angeles County, California, to investigate the correlation between microstructures and large scale structures and to determine the possibility of predicting large scale structures from microstructures.

Methods of petrofabric analysis, such as the determination of dimensional and crystallographic orientation of quartz grains, showed preferred orientations; but no well-defined relationships were noted between micro- and megastructures. The anisotropies observed in the orientation data are attributed to depositional rather than to deformational fabrics.

The study of microfractures, however, provided a means of relating the microscopically observed features to the large structure. Statistically, the microfractures are strongly oriented, and the preferred orientation in any sample can be used to determine the strike and plunge of the anticline within usable angular limits.

INTRODUCTION

Petrofabric analysis is merely a detailed investigation of structural geology and petrology. Most of the field methods of petrofabrics have been used by geologists for years. Much additional information can be obtained, however, by application of laboratory petrofabric techniques. Moreover, if the small scale features of isolated hand specimens can be integrated with the large scale features of major structures, much more structural information can be obtained from cores than is now available.

Van Hise (1896) and Leith (1905) showed that small scale structures, such as drag folds, are related to large scale anticlines and synclines. Similarly, Sander (1930) showed that microscopic folds or crenulations in fine-grained rocks are related to the drag folds. Furthermore, Sander showed that the orientation of rock components, dimensional or crystallographic, may reflect the large structure and indicate the type of movement that caused the structure.

Field geologists engaged in oil exploration have found the concepts outlined by Van Hise and Leith to be useful in deciphering complex structures. Some attempt has been made to use the microscopic counterparts of structural features such as drag folds and fracture cleavage to interpret large structures. Essentially, however, the statistical methods of petrofabrics have been neglected in petroleum geology. Perhaps the fabric approach has not been used because of the lack of completely acceptable theories to account for the observed facts. Nevertheless, the methods and analytic procedures are purely observational; and, as petroleum geologists have never hesitated to make practical applications of empirical observations, the time has come to use the mass of data available from fabric studies.

Most investigations of deformation fabrics have been made on rocks which have undergone severe deformation, that is, metamorphic rocks. Little work has been done on the fabrics of rocks and structures commonly encountered in oil exploration. Metamorphism and structural deformation, however, must be rgarded as progressive processes; therefore, the rock fabrics of slightly deformed structures should be related, although less pronounced, to the rock fabrics of intensely deformed structures. This paper is a discussion of the methods and results of a petrofabric study of a typical California oil-bearing structure, the Pico Anticline.

GEOLOGIC SETTING OF THE PICO STRUCTURE

The Pico Anticline is located about 35 miles northwest of Los Angeles and about 5 miles west of Newhall, Los Angeles County (figs. 1, 2). The area is of historical interest, for the first known California oil production

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was from seeps in Pico Canyon. The first producing well was drilled in Pico Canyon in 1875, and production was made later from wells in Moore, Towsley, Wiley, Rice, and East Canyons.

Stratigraphy

The Pico Anticline lies along the southern margin of the Ventura Basin, which received sediments from the Eocene to the present until an estimated maximum thickness of 44,000 feet accumulated. Three formations are encountered on the Pico structure. They are the Modello (Upper Miocene), Pico (Pliocene), and Saugus (Pliocene).

A complete section of the Modello Formation has not been measured in this area. Kew (1943) reports that 3150 feet of the upper part of the formation is exposed in Towsley Canyon. The formation, as exposed on the anticline, consists of brown, silty, laminated shale with sandstone (arkose) beds and a few conglomerate beds.

The Pico Formation was first defined by Kew (1924), chiefly on the basis of lithology. Later, after examination of foraminifera by Driver and Holman, the contact was placed about 1900 feet lower in the section (Kew, 1943). The Pico overlies the Modello conformably and consists of brown to gray siltstones and silty shales with many lenticular sandstones and conglomerates. The total thickness of the formation as now recognized is about 6700 feet.

The Saugus Formation overlies the Pico unconformably and is composed of slightly indurated, nonmarine sands and conglomerates aggregating 4600 feet near Pico Canyon.

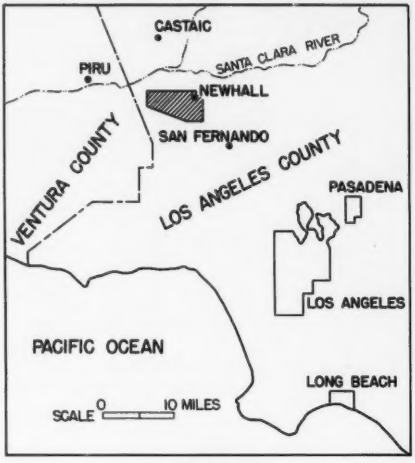

FIG. 1.—Index map.

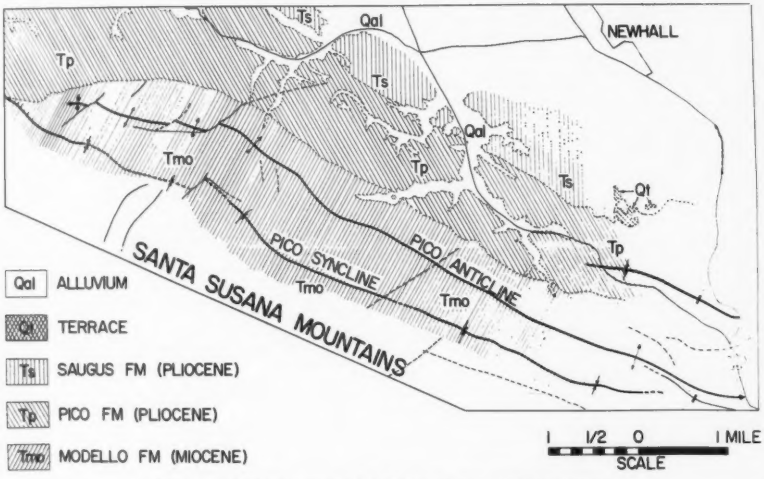

FIG. 2.—Geologic map of vicinity of the Pico Anticline.

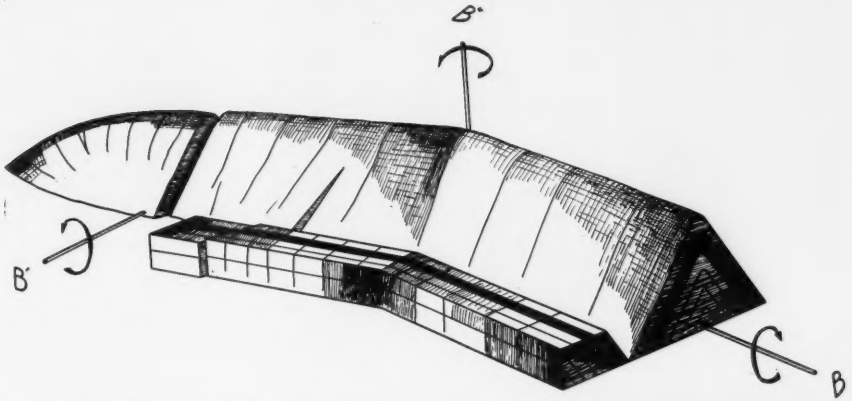

FIG. 3.—Idealized block diagram of the Pico structure showing three axes of external rotation.

Structure

The Pico Anticline is one of the major structural features of the Ventura Basin. It is about 9 miles long and extends from San Fernando Pass westward beyond Pico Canyon (fig. 2). The Pico Syncline parallels the anticline on the south.

The anticline is folded tightly and shows nearly a chevron cross section. The axial plane is nearly vertical and has a strike of about N. 65° W. in the central and eastern part of the structure and about N. 75° W. in the western end of the structure. Dips on the flanks of both structures are commonly over 50 degrees, and some overturned beds occur on the north limb of the anticline. The anticline plunges on both the east and west ends; however, the closure on the western end is easier to observe because of better outcrops.

Several normal faults with maximum throws up to 500 feet cut the structure axes. Most of these faults have a northeast-southwest strike.

RESULTS OF THE FIELD INVESTIGATION

An intensive field study was made of the Pico Anticline in the Pico Canyon area. The remainder of the anticline and the adjacent Pico Syncline were studied in less detail. The field investigation consisted of a study of the nature of the deformation, assignment of fabric axes, measurements of s-surfaces, and collection of oriented samples.

Nature of the Deformation

The Pico Anticline and Syncline are relatively simple flexure slip folds. The sandstone and conglomerate beds behaved as competent layers during folding. They were mostly deformed by laminar slip on bedding planes rather than by plastic flow; consequently, they show little stretching on the flanks or thickening on the axes of the folds. In a few places near the tightly compressed anticline axis, thin sandstone beds were drag-folded and sheared. The shales, however, behaved as incompetent members and were deformed plastically; therefore, the beds were thickened on the axis and thinned on the flanks of both folds. Some outcrops were noted on the flanks of the anticline in which shale beds were pinched out completely. In many shale beds, individual layers were crumpled and entire blocks were rotated to such an extent that the bedding was obliterated.

Assignment of Fabric Axes

For convenience of reference, conventional fabric axes were assigned to the Pico structure. The *a* directions are normal to the axial planes of the folds, the *b* directions are parallel to the fold axes, and the *c* axes are vertical. The external rotation of bedding planes about the axes of the folds is the basis for terming the structure a B-tectonite. Field study showed that three mutually perpendicular axes of rotation occur on

the anticline, thus the Pico Anticline may be classified as a $B \perp B' \perp B''$ tectonite.

Figure 3 shows the mutual relations of the three B axes. One B axis coincides with the axis of the anticline. A second axis of rotation is apparent on the anticline; that is, the axis about which the beds rotate to form the plunging nose. This axis, B′, is horizontal and normal to the axial plane of the folds. A third axis, B″ is indicated by the change in strike of the anticline axis near Big Moore Canyon. Although the rotation about this vertical axis is slight (10° to 15°), the effect is evident in some of the fabric diagrams presented later.

S-Surfaces

Planes of stratification are the most prominent s-surfaces in nearly all of the sandstones. In a few outcrops of massive sandstone, however, careful observation is necessary to distinguish between planes of stratification and *ac* joints. In the shales, joints are often the dominant s-planes, and in many outcrops the s-plane of stratification cannot be detected.

The dips of the bedding on both limbs of the anticline and syncline are mostly from 50° to 90°, and some of the beds on the north limb of the anticline are overturned. The bedding is nearly horizontal along most of the length of the anticline axis, but noticeable plunge begins near the top of the ridge just east of the Pico Oil Field and increases to a maximum of about 40° at the nose of the fold. Dip and strike measurements are difficult to obtain along the axis of the syncline because the trough is occupied by contorted shale, but the strike of nearby sandstone beds indicates that the axis is nearly horizontal.

Faults form a second set of s-surfaces. A few major faults cut the axis of the anticline at angles of 30° to 40°. Most of them strike northeast, although a few strike northwest (fig. 2). These faults are s-planes of shear oriented symmetrically to the *a* fabric direction of the fold. Minor reverse faults with displacements of a few inches to a few feet are found in all parts of the structure. Most of the minor faults are associated with joint systems.

The joint sets are additional s-planes. They range from close-spaced, discontinuous fractures less than an inch apart to wide-spaced, continuous fractures. The jointing is influenced by thickness of the beds and position on the fold. The spacing of the joints approximately corresponds to the thickness of the individual beds; the joints are close-spaced in the thinly bedded shales and silty sands and wide-spaced in thick sandstone beds. Jointing is most intense near the axis of the anticline and around the plunging nose, but it is less severe on the limbs of the fold.

The joint pattern of the Pico structure is shown on the elemental fabric diagram in figure 4A. The incomplete girdle in the diagram corresponds to preferred joint orientations with strikes ranging from N. 10° E. to N. 50° E. and with dips ranging from 90° to 40° S. E. Diagrams prepared for different parts of the anticline show that the preferential orientation of joints in the central part to the fold is vertical and perpendicular to the axial plane, whereas joints on the plunging nose are perpendicular to the axial plane but dip eastward in inverse proportion to the amount of plunge of the fold to the west. Therefore, the preferred orientation of joints is in the *ac* fabric plane of the fold; that is, normal to the B fold axis. The change in dip of the joints reflects external rotation about the B′ axis.

Joints of this type are attributed to tension acting perpendicular to the direction of compression during deformation. Failure in this manner implies that the overburden was relatively thin at the time of compression so that lateral relief by rupture was easy.

The joint diagram (fig. 4A) also contains several minor joint pole maxima. These minor maxima correlate with the shear directions of faulting described above. The joint pole diagrams have minimum concentrations in areas corresponding to joints parallel to the fold axis; that is, no fracture cleavage or tension fractures occur parallel to the fold axis. This is additional evidence that the rocks were not buried deeply during deformation.

Collection of Specimens

Over 200 samples from the Pico structure were selected for laboratory work. These were carefully oriented and marked before removal from the outcrop. Most of the

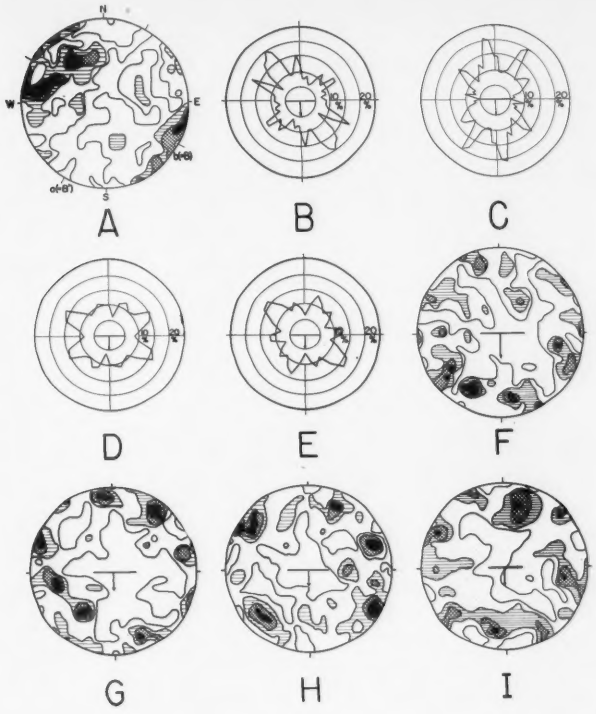

FIG. 4

Diagram	*Type of Diagram*	*Location*	*Orientation of Section*	*Number of Points*	*Contours—Percent*
A	Poles to joint sets	Pico Anticline		210	>4, 3, 2, 1
B	Apparent elongation of quartz grains	Pico Anticline	Parallel to bedding (*ab* section)	250	
C	Apparent elongation of quartz grains	Pico Anticline	Parallel to bedding (*ab* section)	250	
D	Apparent elongation of quartz grains	Near San Juan Capistrano	Parallel to bedding (*ab* section)	200	
E	Apparent elongation of quartz grains	Near San Juan Capistrano	Parallel to bedding (*ab* section)	200	
F	Quartz crystallographic axis orientation	Pico Anticline	Parallel to bedding (*ab* section)	200	>4, 3, 2, 1
G	Quartz crystallographic axis orientation	Pico Anticline	Parallel to bedding (*ab* section)	200	>4, 3, 2, 1
H	Quartz crystallographic axis orientation	Near San Juan Capistrano	Parallel to bedding (*ab* section)	200	>6½, 4, 2, 1
I	Quartz crystallographic axis orientation	Near San Juan Capistrano	Parallel to bedding (*ab* section)	200	>4, 3, 2, 1

samples collected were sandstones because they are easily studied by optical methods of petrofabrics. As a control to distinguish deformational from nondeformational fabrics, several oriented samples were collected near San Juan Capistrano, California. These samples are similar in age and lithology to those from the Pico Anticline but have undergone only gentle structural upwarping.

RESULTS OF THE LABORATORY INVESTIGATION

The purpose of the laboratory investigation was to determine which small-scale features of the oriented rock specimens could best be correlated with the large-scale features observed in the field.

Initially, three oriented thin sections were cut from each sample. One section was cut parallel to the bedding (*ab* fabric plane), a second was cut perpendicular to the bedding and parallel to the dip, and a third was cut perpendicular to the bedding and parallel to the strike. Later, sections cut parallel to the bedding were found sufficient for the desired work, providing an occasional section was cut perpendicular to the bedding as a check.

Petrography of the Rocks

Most of the samples studied in the laboratory were from the Modello (Miocene) Formation, although a few samples from the lower part of the Pico (Pliocene) Formation were included.

Most of the sandstones may be classified as arkoses. The colors range from light greenish or yellowish gray to dark yellowish brown. The constituent grains are: quartz, 50 to 70%; feldspar, 10 to 40%; clastic biotite, magnetite, and hornblende, 5%; and minor amounts of rock fragments and zircon. The cementing material is clay, carbonate, iron oxide, or any combination of the three. The samples cemented with carbonate are very well indurated, whereas most of those cemented with clay or iron oxide are friable and porous. Secondary minerals include calcite, dolomite, siderite, sericite, chlorite, kaolinite, gypsum, and iron oxides.

The grains are mostly angular to subround, and the size sorting is fair in most samples. The grain sizes range from a fine pebble conglomerate to a very fine sand or even silt. Quartz and feldspar grains are slightly inequant, and the longest dimension of most grains occurs in the bedding plane.

Bedding is shown best by the subparallel orientation of planar and linear minerals such as biotite and hornblende. Variation in grain size or sorting may define the bedding, but in some homogeneous samples the bedding planes cannot be detected by visual observation.

Many samples show slickensided surfaces along bedding planes, small faults and fractures cutting the bedding, and zones of crushed grains or contorted platy minerals. The individual grains, particularly quartz, show deformation features, such as undulatory extinction and several types of fracture. A preferred orientation of grain fractures was noted on polished surfaces of a few samples. The microfractures are described in detail in a later section.

Quartz Dimensional Orientation

In many structures, deformation by componental movement or solid flow results in a preferred dimensional orientation of fabric elements such as fossils, ooids, concretions, pebbles, or mineral grains. In most folds these fabric elements are elongated parallel to the fold axis (B direction).

Two types of dimensional orientation are characteristic of B tectonites, one in which the orientation results from deformation of the fabric element and another in which the orientation results from rotation of originally inequant elements to a parallel position. In the first type, fabric elements are elongated and oriented by rolling in the rock during folding, much as a ball of clay is elongated perpendicular to the direction of movement when rolled by the hand on a flat surface. In the second type, the elements are elongate before deformation, and the movement rotates the elements to positions perpendicular to the movement direction.

In some structures the fabric elements are oriented with the long axes perpendicular to the fold axis and parallel to the *a* direction. This is attributed to deformation of the elements by stretching parallel to the movement direction. Elongated grains on slickensided surfaces are examples of this process.

The dimensional orientation of grains in selected planes may be measured in thin sections. However, sections provide only a

means of measuring the apparent elongation; they do not necessarily contain the actual axis of elongation. Nevertheless, in structures like the Pico Anticline, sections parallel to the bedding are in the *ab* plane; therefore, bedding plane sections are best for measuring dimensional orientation.

The distributions of apparent elongation axes in the bedding planes of several sandstone samples from the Pico Anticline are shown in figures 4B and 4C. The maxima in these and other diagrams, which are not shown, are not strong. In a few samples the preferred orientation can be correlated with the fold axis or direction of movement, but in most samples the relationship between dimensional orientation and deformation is not clear.

Nevertheless, a preferred dimensional orientation does exist in many of the diagrams, and the plots are bilaterally symmetrical. Elongation diagrams prepared from undeformed rocks (figs. 4D and 4E) are similar to those prepared from the Pico rocks. The symmetry and preferred orientation of the Pico samples, therefore, may be partly a relic dimensional fabric formed when the sands were deposited.

Quartz Crystallographic Orientation

The fracture hypothesis, as yet not completely proved, is used by some structural petrologists to explain quartz orientation in tectonites (Fairbairn, 1949). Briefly, this hypothesis is based on the assumption that upon initial deformation, quartz grains are fractured into elongate fragments parallel to certain crystallographic directions, that is, the vertical, horizontal, or rhombohedral edges of a crystal. During continued deformation, the needles of quartz are rotated so that their elongation axes are parallel to the direction of movement. Later recrystallization may destroy the elongated shape and leave only a preferred orientation of the crystallographic axes.

If this hypothesis is to be accepted; then, depending upon the degree of deformation, a structure should contain one of the following:

1.—Needle-like fragments of quartz.
2.—Dimensional orientation of elongate quartz fragments.
3.—Recrystallized quartz with a preferred crystallographic orientation related to the structure.

Thin sections of sandstones from the Pico structure were examined carefully for all of these criteria, but none was observed. The quartz grains are fractured extensively (see the section on microfractures), but few fragments are needle-shaped. Furthermore, the quartz grains do not show a strong dimensional orientation related to the deformation (see section on dimensional orientation). Finally, little quartz recrystallization is visible in any of the sections. Deformation may have caused a slight etching or solution of quartz grains, but only slight secondary growth is noticeable on the quartz grains.

A second hypothesis of quartz orientation in tectonites is based upon translation-gliding in the crystal lattice (Fairbairn, 1949). This hypothesis also, is unproved experimentally. The evidence for orientation rests upon the observation of deformation lamellae in quartz grains. The lamellae are bright, narrow lines in the quartz grains and are comparable to the glide lines observed in many metals (Riley, 1947).

Translation-gliding cannot be accepted as the method of quartz deformation in the Pico Anticline because no indigenous deformation lamellae were observed.

A number of quartz crystallographic fabric diagrams were prepared from thin sections cut normal to the *b* and *c* fabric axes of sandstones from the Pico Anticline. Representative diagrams from sections cut normal to fabric *c* are shown in figures 4F and 4G.

The preferred orientation of quartz axes as shown by the maxima on the diagrams is not strong. In order to determine whether or not the observed pattern could be the result of random distribution of quartz axes, a standard statistical test, the Chi-square test (Fairbairn, 1949), was applied to several of the diagrams. The results indicate that an assumption of isotropy is not justified.

Granted that the quartz crystallographic distribution is anisotropic, can the observed fabric be correlated with the deformation? The fracture hypothesis of quartz orientation in tectonites postulates the location of certain maxima in fabric diagrams. A number of investigators report in the literature

that certain of these maxima are actually present in tectonite fabrics. Most of the rocks they studied, however, were severely deformed by plastic flow. Little work has been done on quartz orientation in large folds like the Pico structure in which little plastic flow occurred. Attempts to relate the maxima in the Pico fabric diagrams to those postulated by the fracture hypothesis or reported in the literature were unsuccessful. As noted above, the translation-gliding hypothesis also is untenable because of the absence of deformation lamellae.

The possibility remains that the observed quartz fabric is not deformation but depositional. Certain features of the fabric diagrams provide evidence to support this possibility. Measurements were made to determine the relation between optic axis and elongation axis in individual loose quartz grains from the Modello Sandstone. The results are shown in figure 6. This graph shows that the long axes of the grains are preferentially oriented parallel to the rhombohedral direction in quartz 38° from the crystallographic C axis. This observation also was reported by Ingerson and Ramisch (1942). Therefore, if the grains were deposited with their long axes parallel to the bedding but in random direction in the bedding plane, their crystallographic axes would be represented by a band or girdle about 38° wide around the periphery of an *ab* fabric diagram. If, in addition, the elongated grains were oriented directionally in the bedding plane by the depositing medium, their crystallographic axes would be represented by isolated maxima within 38° of the bedding plane of the *ab* diagram.

The majority of the maxima occur within 35° of the bedding plane in all *ab* diagrams prepared from Pico rocks, which suggests a depositional origin of the observed fabric.

Further evidence that the observed quartz orientation fabric is primarily depositional is afforded by consideration of the symmetry of the diagrams. A plane of monoclinic symmetry may be observed in most of the diagrams. Such symmetry is attributed to unimpeded flow (Fairbairn, 1949). The symmetry may be deformational, as the result of plastic flow of the rock, or depositional, as the result of fluid flow of the depositing medium. If the symmetry is due to deformation, the symmetry planes should be parallel to the *ac* fabric planes of the structure. In a few samples the symmetry plane does occupy the *ac* fabric plane, but in most samples it does not. Fabric symmetry, therefore, is not a reliable indicator of deformation axes in this structure.

Finally, evidence that the quartz fabric is depositional is provided by a comparison of the Pico fabric with the fabric of undeformed rocks. Quartz orientation diagrams were prepared from the samples of undeformed sandstones collected near Capistrano, California (figs. 4H and 4I). The characteristics of peripheral maxima and monoclinic symmetry are similar in fabric diagrams of both rock suites.

Microfracture Orientation

Well-developed systems of fractures in individual quartz grains (microfractures) were observed in the examination of polished surfaces and thin sections. Because the macrofractures or joints have an orientation closely related to the structure, the orientation of the microfractures was studied to determine whether they are also related to the structure. This phase of the fabric study proved to be the most useful in correlating microscopic features with the deformation picture.

The only microfractures plotted were those in individual quartz grains. Fractures in feldspar were not used because the feldspars tend to fracture parallel to cleavage planes. The quartz fractures are of several types as follows:

1.—Clean, straight or slightly curved fractures crossing one or more grains. May or may not show displacement.
2.—Relatively large, straight or slightly curved fractures filled with carbonate, sericite, or clay.
3.—Planes of bubbles or liquid inclusions ranging from almost continuous sheets to separate round bubbles. May or may not cross an entire grain.

Studies were made to determine whether or not any one of these types was more significant in the fabric analysis than the others, but no distinction could be made.

Some fractures of the first type may be formed in the rock during the preparation of thin sections, but even the fabric diagrams prepared from sections suspected to

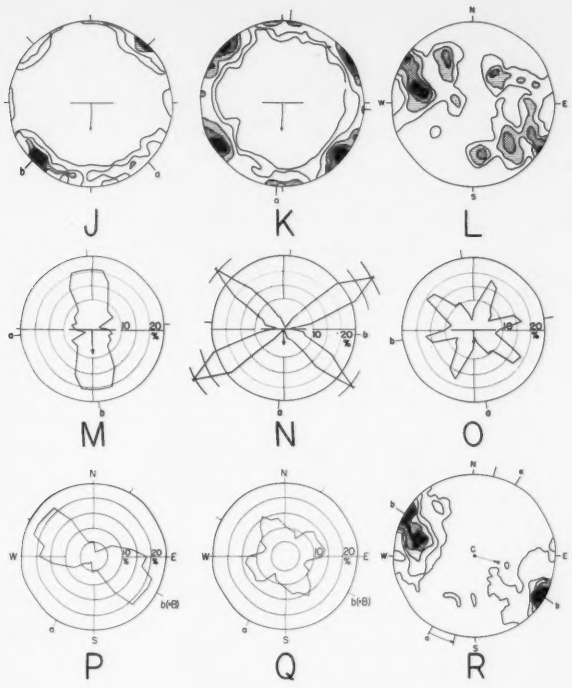

FIG. 5

Diagram	*Type of Diagram*	*Location*	*Orientation of Section*	*Number of Points*	*Contours—Percent*
J	Poles to quartz microfractures (Universal stage)	Pico Anticline	Parallel to bedding (*ab* section)	200	>10, 8, 6, 4, 2
K	Poles to quartz microfractures (Universal stage)	Pico Anticline	Parallel to bedding (*ab* section)	200	>8, 6, 4, 2, 1
L	Composite diagram from 22 quartz microfracture diagrams			43	>10, 7, 4, 2
M	Poles to quartz fractures (two-dimensional)	Pico Anticline	Parallel to bedding (*ab* section)	200	
N	Poles to quartz fractures (two-dimensional)	Pico Anticline	Parallel to bedding (*ab* section)	100	
O	Poles to quartz fractures (two-dimensional)	Pico Anticline	Parallel to bedding (*ab* section)	100	
P	Composite diagram of major microfracture maxima from two-dimensional diagrams	Pico Anticline		184	
Q	Composite diagram of minor microfracture maxima from two-dimensional diagrams	Pico Anticline		336	
R	Composite diagram of microfracture maxima from all two-dimensional diagrams	Pico Anticline		180	>8, 6, 5, 3, 1

contain a number of induced fractures contain preferred orientations closely related to the structure. Therefore, significant errors are not caused by fractures induced during the cutting of the rocks. In fact, the orientation of induced fractures may be partly dependent upon strain in the quartz grains inherited from the deformation accompanying folding.

Fractures of the third type consisting of planes of inclusions may be partly healed fractures of the first type (Tuttle, 1949).

A suite of 28 sandstone specimens was selected, and oriented thin sections were prepared from each sample. These rocks were chosen from different parts of the structure, and they range in lithology from coarse siltstones to pebble conglomerates.

The universal stage was used to prepare fabric diagrams from each section to show the distribution of poles (normals) to microfracture planes. All diagrams from sections cut parallel to the bedding (*ab* plane) contain strong maxima on the periphery (figs. 5J and 5K).

A good correlation exists between the microfracture maxima and the fabric axes of the folds. In diagrams with a single maximum (fig. 5J), the microfractures are oriented preferentially in the *ac* fabric plane; i.e., normal to *b* (=B). In diagrams with two maxima (fig. 5K), the microfractures are oriented preferentially in two planes bisected by the *ac* fabric plane.

The spatial distribution of the planes of maximum microfracturing is shown in figure 7. In the central part of the anticline and all along the syncline, the fold axes are approximately horizontal; i.e., the *ac* fabric plane is vertical and perpendicular to the fold axis. The microfractures in these parts of the structure are preferentially oriented vertical and normal to the fold axis (parallel to the *ac* plane), or they are preferentially oriented in two vertical planes bisected by the *ac* fabric plane.

The planes of microfractures on the plunging west nose of the anticline show the same relation to the fabric axis, and they dip to the east as the *b* (=B) axis plunges to the west. That is, the microfractures rotate about B′.

The correlation between microfracture orientation and the large structure is shown also by the composite diagrams of figure 5L. In figure 5L, the maximum on the periphery represents the concentration of vertical *ac* microfractures in the central part of the structure; whereas, the maximum about 30° from the margin represents the east-dipping microfractures on the plunging nose of the anticline. The two maxima are connected by an incomplete girdle which indicates the extent of rotation about B′. The *ac* plane is a good plane of symmetry.

Preparation of fabric diagrams with the universal stage is time-consuming and tedious, and a more rapid method of analysis is desirable if large numbers of samples are studied. As noted above, the three-dimensional diagrams show that the microfractures significant in structure analysis are *ac* fractures oriented perpendicular to the bedding; therefore, a two-dimensional plotting method is justified for determining the *a* direction in the bedding plane.

Two-dimensional plots of microfracture orientation were prepared from 210 sections cut parallel to the bedding. A few representative diagrams are shown in figures 5M, 5N, and 5O.

Each diagram may be classified as one of three types containing: (1) a single strong maximum (fig. 5M), (2) a pair of maxima about equally developed, (fig. 5N), or (3) combinations of (1) and (2) symmetrically arranged (fig. 5O).

The structure-microfracture relations also are shown by composite diagrams. The major microfracture maxima were selected from each two-dimensional diagram and combined in figure 5P. This diagram clearly shows that the major direction of fracturing is normal to the axis of the structure; i.e., in the *ac* plane.

The minor microfracture maxima were combined in figure 5Q. This diagram shows two directions of fracturing corresponding to the shear directions resulting from compression in the *a* direction.

As noted above, the diagrams prepared with the universal stage indicate that the microfractures are preferentially oriented perpendicular to the bedding. Therefore, if the dip and strike of the bedding are known, the attitude of the *ac* fracture fabric plane can be determined from the two-dimensional plots.

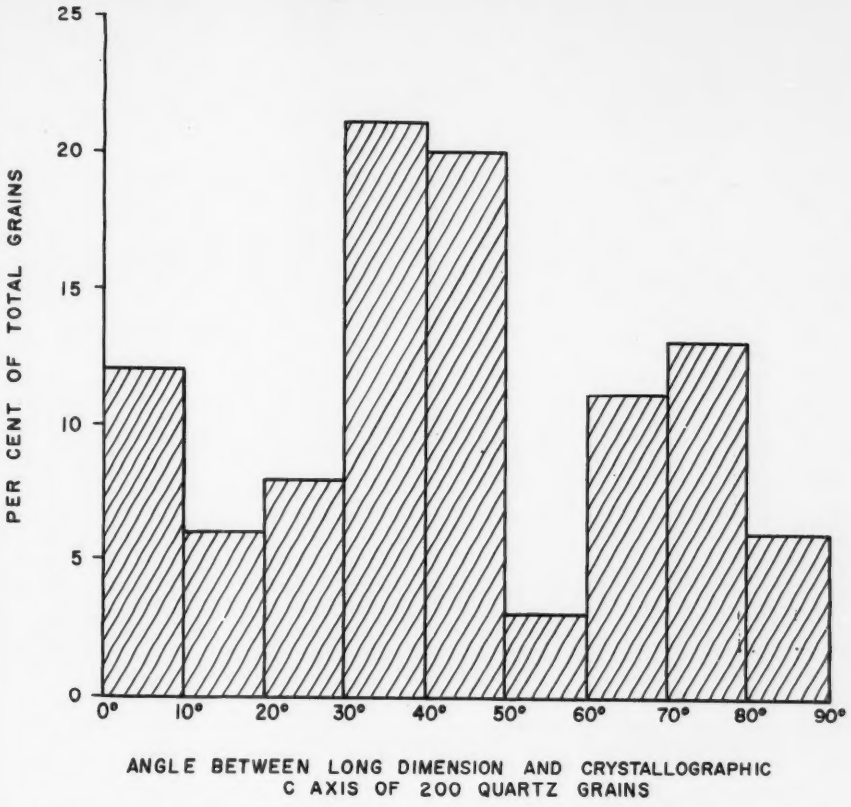

FIG. 6.—Relation between long dimension and crystallographic C-axis of 200 quartz grains.

Figure 5R is a composite diagram prepared from all two-dimensional microfracture plots. The elongated maximum forms part of a girdle showing the change in strike of the structure from N. 65° W. to N. 80° W. (rotation about B″) and a change in the axis from horizontal to about 25° W. plunge (rotation about B′).

The quartz contains some fractures not visible in sections cut parallel to the bedding. Sections perpendicular to the bedding contain a microfracture set oriented parallel to the bedding. In a few sections, pairs of microfracture maxima are equally inclined to the bedding plane. Most of the fractures parallel to or at low angles to the bedding are attributed to failure in tension or shear as a result of laminar slip parallel to the bedding planes. These fractures are not consistent enough to be used for structural analysis.

Structure Prediction

If the composite microfracture diagrams are closely related to the major structure, how much information can be obtained from one or a limited number of oriented samples? It was noted in the two-dimensional microfracture diagrams that the angle between dip direction and microfracture direction ranges from 0° for samples from the flanks of the fold to 90° for samples from the plunging nose. This relationship provides a method of obtaining large-struc-

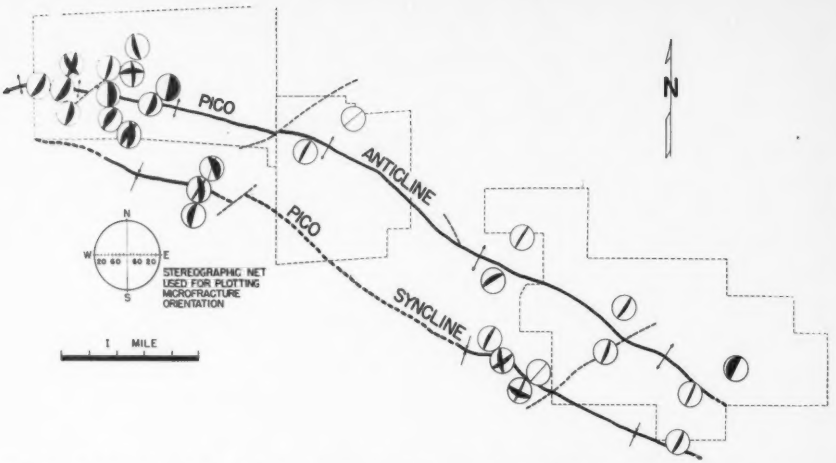

FIG. 7.—Preferred spatial orientation of microfractures as determined by means of the universal stage.

ture data from a microfracture study of a few samples.

The strikes and dips of the *ac* fracture fabric planes were determined from the two-dimensional plots of all samples from the Pico Anticline by the method described above. Figure 8 shows the change in the angle ϕ between the direction of dip of bedding planes and the strike of the *ac* fracture fabric plane as the direction of dip of the bedding changes at various positions on the fold.

The scatter on the diagram is due to several factors which include:

1.—Change in strike of the anticlinal axis along its length (B″ rotation).
2.—Rotation of beds due to minor crumpling, cross folding, and slump.
3.—Experimental error introduced during collection, cutting, and grinding of the rocks or in measuring the fracture orientations.

The statistical data from figure 8 are tabulated in table 1. The calculated regression line has a slope near the value of 1; thus, a direct relationship exists between dip direction and the angle ϕ. The correlation coefficients calculated for the observed data are near unity (table 1); therefore, the angle ϕ is approximately equal to the angle between the strike of the bedding and the axis of the anticline. The frequency distribution diagram (fig. 9) shows that the strike of the fold axis is indicated within 20° by about 80 percent of the samples.

An additional test of the statistical data is provided by the close correspondence between the strike of the fold axis as determined on the regression line and the strike of the fold as observed in the field. The strike of the fold axis is indicated on the regression line by the point at which the angle ϕ equals 90°.

On the basis of this study, it may be concluded that the relation between the spatial orientation of microfractures and the attitude of bedding serves to locate a sample on the Pico Anticline with respect to the axis

TABLE 1.—*Statistical analysis of microfracture data*

Slope of calculated regression line	0.992
Correlation coefficient	.970
Strike of fold axis observed in the field	N. 70° W.
Strike of fold axis computed from microfracture data	N. 68° W.
Percent of observed measurements falling within 20° of the regression line	82%
Percent of observed measurements falling within 15° of the regression line	70%
Percent of observed measurements falling within 10° of the regression line	51%

of the fold. Because B is normal to the *ac* plane, the strike of the anticline is the direction normal to the strike of the *ac* plane, the plunge of the fold is equal to the complement of the dip (hade) of the *ac* plane, and the direction of plunge is opposite to the direction of dip of the *ac* plane. The analysis, however, does not indicate the distance from the sample to the axial plane or crest of the anticline. Whether these relationships are true of other structures must be determined by additional petrofabric analysis.

SUMMARY AND CONCLUSIONS

The Pico Anticline and adjacent Pico Syncline are relatively simple folds in which most of the deformation occurred by laminar slip along bedding planes. The sandstones and conglomerates, although most of them are poorly indurated, behaved as competent layers; thus, they show little evidence of solid flow. The shales, on the other hand, are severely contorted and deformed by plastic flow.

The fold is classified as a $B \perp B' \perp B''$ tectonite. B is the axis of the anticline, B′ is the axis of rotation associated with plunge, and B″ is the axis of rotation associated with a change in strike of the fold.

Bedding planes, faults, and joints are the prominent s-surfaces visible in the field. Most of the larger faults are attributed to shear resulting from compression normal to the fold axis. The joints are oriented statistically in the *ac* plane of the fold. They are attributed to tension acting normal to the compression direction.

Methods of petrofabric analysis were used in the laboratory studies in an attempt to correlate micro- with macrostructure. The quartz of the Pico structure has both preferred dimensional and crystallographic ori-

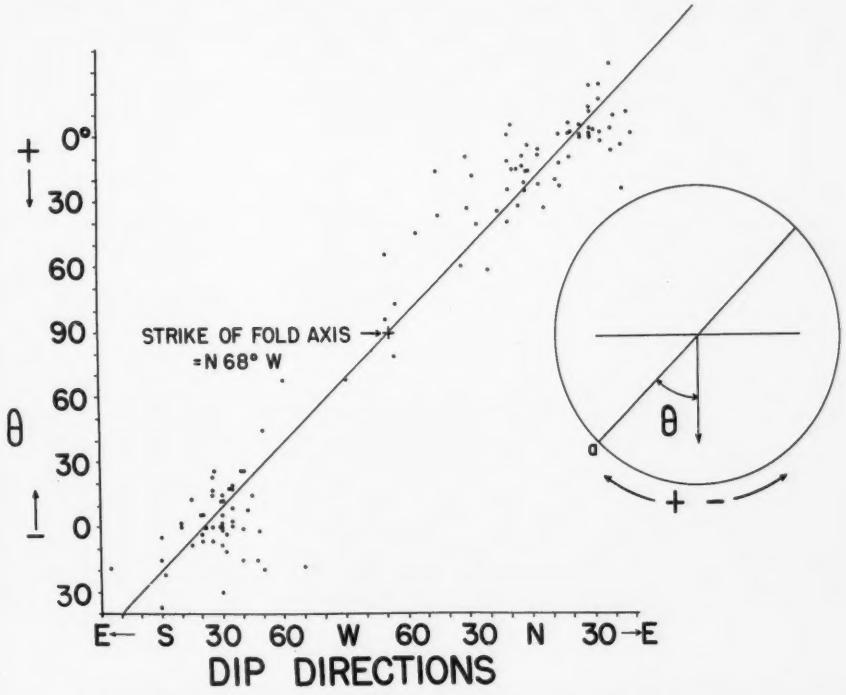

FIG. 8.—Relationships between dip direction and microfracture-dip angle. ϕ equals the angle between the dip direction and the strike of the *ac* microfracture plane.

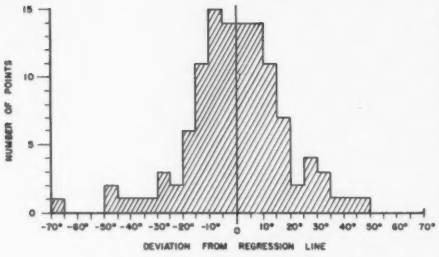

FIG. 9.—Frequency distribution of points from figure 8.

entation, but the quartz fabric cannot be correlated closely with the deformation accompanying folding. The evidence suggests that the fabric of the quartz is mostly an inherited depositional fabric.

The study of microfractures provides the best correlation between microscopically observed features and the large structure in this area. The microfractures show a preferred orientation in the *ac* plane of the fold or in two sets bisected by the *ac* plane. Moreover, the plunge of the fold is related to the plunge of the normal to the plane of microfractures. The preferred microfracture orientation can be obtained by a simple two-dimensional technique. Spatial orientations can be obtained from the two-dimensional diagrams without the use of a universal stage.

The practical significance of the microfracture orientation is that the preferred orientation of microfractures in a sample indicates the trend and plunge of the fold from which the sample was obtained. From about 80 percent of the samples studied, the strike of the anticline axis can be determined within 20°. The direction and amount of plunge of the fold in the vicinity of the sample can be determined with about equal accuracy.

In conclusion, this study provides evidence that even in relatively slightly deformed rocks the microfabric can be correlated with the major structure. Conventional and new methods of petrofabrics provide much more structural information from outcrop samples and drill cores than the customary direction and amount of dip. This information should be particularly valuable in the exploratory drilling of structures in areas of limited geological knowledge. However, the techniques and interpretations described in this report should not be hastily or indiscriminately applied. Only one rock type in one structure has been investigated. After a number of structures have been studied it may be possible to outline petrofabric techniques to solve some of the structural problems associated with petroleum exploration.

ACKNOWLEDGMENTS

A number of people aided in this petrofabric study, and their help is gratefully acknowledged. They include Messrs. W. F. Barbat, E. C. H. Lammers, H. L. Driver, J. P. Bailey, and W. H. Holman of the Standard Oil Company of California, and Messrs. N. A. Riley and D. H. Davis of the California Research Corporation.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 27, NO. 3, PP. 265–270
FIG. 1, SEPTEMBER, 1957

AUTHIGENIC BROOKITE IN CRETACEOUS GALLUP SANDSTONE, GALLUP, NEW MEXICO[1]

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ABSTRACT

Brookite-bearing sandstone near Gallup, New Mexico, was found by a Navajo Indian while prospecting for uranium. Owing to its high content of zircon the radioactivity of the sandstone is about 3 to 5 times above background. Though it was first considered as "uranium ore" by the prospector, the sandstone was referred to locally as the "Gallup titanium deposit" after titania minerals were found to be the important constituents of the sandstone.

The light minerals of the sandstone include albite, orthoclase, and quartz. The heavy minerals include anatase, apatite, brookite, ilmenite, magnetite, rutile, spessartite, sphene, tourmaline, and zircon. The clay minerals include kaolinite, montmorillonite, and limonite.

Leucoxene has formed at the expense of ilmenite. Authigenic brookite, which occurs mostly as small boxworks between other detrital mineral grains, formed at the expense of both ilmenite and leucoxene. Detrital rutile was not positively identified.

INTRODUCTION

Locality.—Though the Cretaceous Gallup Sandstone crops out in many places in the San Juan Basin in northwestern New Mexico, the lens of rich brookite-bearing rock at the top of one member of the Gallup Sandstone discussed in this paper crops out in SE¼ sec. 32, T. 15 N., R. 19 W. It is about 1¾ miles by truck trail from Highway 66 and the A.T.&S.F. Railroad tracks. It is about 6 miles from Gallup and is about 400 feet in altitude above the highway and railroad

Discovery.—Late in 1955 a Navajo Indian discovered the brookite-bearing sandstone while prospecting for uranium with a scintillation counter. Owing to its high content of zircon, the radioactivity of the sandstone is about 3 to 5 times above background. The sandstone was considered as "uranium ore" by the prospector until titania minerals were identified to be the chief constituents of the sandstone in the laboratory of the State Bureau of Mines and Mineral Resources. Subsequently it was referred to locally as the "Gallup titanium deposit."

The sandstone forms a gently sloping bench about 1000 feet wide above a 70-foot cliff and below a 50-foot slope of coal and carbonaceous shale. The bench is cut by two narrow canyons which divide the sandstone outcrop into three parts.

[1] Published with the permission of the Director, New Mexico Bureau of Mines and Mineral Resources. Manuscript received August 30, 1956.

GEOLOGY

The brookite-bearing sandstone is at the top of the middle sandstone unit of the Gallup Sandstone, just below the Myers or Richards Coal of Sears (1925). It lies one quarter mile east of the axis of the Torrivio Anticline, where the bedded rocks of the Mesaverde Group dip about 3 degrees to the east-northeast.

The brookite-bearing sandstone has a characteristic olive-greenish tinge (G.S.A. color chart No. 10Y6/2) with irregular reddish streaks. It is markedly heavier than the associated sandstone. These features, together with its radioactivity, derived from the uranium-bearing zircon, permitted its being traced along a N. 25° W. direction for about 1000 feet. The width of the sandstone lens ranges from 100 to perhaps 200 feet or more. Owing to lack of any vertical exposures, its maximum thickness could not be determined but appeared to be at least one foot. In a few places it could be as much as 5 feet thick. Apparently this brookite-bearing sandstone occurs as an elongated lens. Outcrops of sandstone of similar mineralogy in a number of places in the San Juan Basin have been reported as titaniferous black sand deposits by Chenoweth (1956).

The trend of the Cretaceous shoreline at the time of deposition of this sandstone is almost identical with that of the brookite-bearing lens (Sears, Hunt, and Hendricks, 1941, pl. 26, line 2–2′). Chenoweth (1956) regards the Gallup Sandstone as a regressive

Upper Cretaceous littoral marine sandstone. The brookite-bearing sandstone is believed to be a titaniferous beach placer (Allen, 1956) and analogous to modern chromite sands on the Oregon coast (Griggs, 1945). Bailey and his co-workers (1956) have pointed out that the occurrence of ilmenite sands, either as modern beach sands or as consolidated sandstone such as the Gallup Sandstone, is rather common.

MINERALOGY

The following minerals of the brookite-bearing sandstone have been identified optically by immersion and by X-ray powder diffraction methods, differential thermal analysis, and by use of the electron microscope.

Quartz.—The largest detrital grains in the brookite-bearing sandstone are quartz which is generally opaque white or colorless. Large grains are sub-rounded to sub-angular, and smaller grains are angular to sub-angular. The size ranges from 0.1 to 0.4 mm in diameter. Stratification of the sandstone is revealed by the distribution of quartz grains (fig. 1).

Albite.—The grains are sub-angular, and about 0.2 mm in diameter. Albite twinning lamellae are distinct in most of the albite grains.

Orthoclase.—Carlsbad twinning was not observed in the few grains of orthoclase. Not enough feldspar grains were found to determine whether or not they are authigenic according to the criteria used by Baskin (1956).

Kaolinite.—X-ray powder diffraction patterns show that kaolinite is the predominant clay mineral. An extensive study of the clay fraction of the Cretaceous sediments of the San Juan basin by Max Willard of the New Mexico Bureau of Mines and Mineral Resources (Oral communication) showed that the dominant clay mineral in the Gallup Sandstone is kaolinite, whereas montmorillonite is second in abundance. Halloysite and illite possibly are present in small amounts.

Montmorillonite.—Montmorillonite was

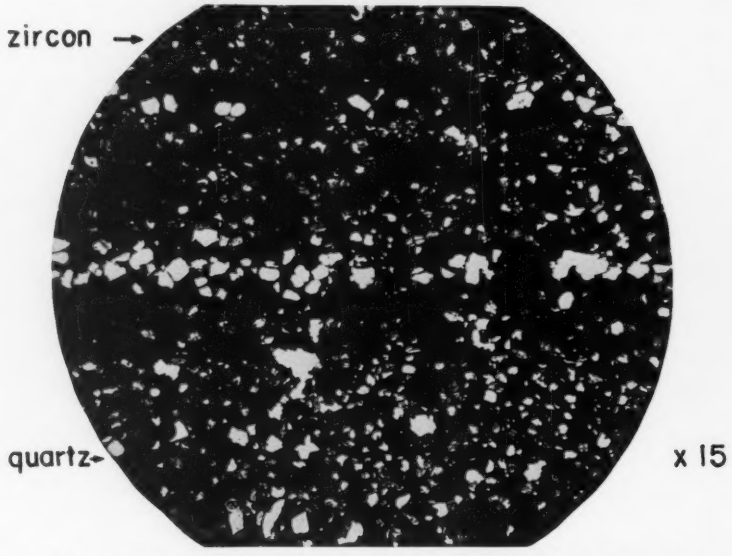

FIG. 1.—Photomicrograph of brookite-bearing sandstone. Stratification of the sandstone is shown by the arrangement of quartz grains. The clear grains are quartz. The cloudy grains are zircon. Opaque titania minerals and the altered ilmenite are hidden in the black background, which is black to brownish-yellow in thin section.

identified by X-ray powder diffraction method and by an electron micrograph.

Limonite.—Limonite is present in the clay fraction.

Anatase.—Only one dark brown partly pyramidal grain of anatase was identified. It has extreme birefringence and is optically uniaxial negative.

Apatite.—A few small prismatic crystals were found.

Brookite.—Most of the brownish-yellow brookite occurs as small boxworks between other detrital mineral grains, but some of it occurs as small plates in leucoxene, or attached to some altered ilmenite. Some brookite was lost when the clay fraction of the sandstone was removed by elutriation. Small boxworks of brookite were collected by using a small tweezer and were identified by X-ray powder diffraction method.

Ilmenite.—Slightly magnetic grains of ilmenite are iron black and have metallic luster. The size ranges from 0.1 to 0.2 mm in diameter. X-ray powder diffraction pattern shows that these grains are predominantly rutile with a small percentage of ilmenite. Leucoxene is attached to some of the iron black grains.

Magnetite.—A few highly magnetic octahedral crystals were found.

Rutile.—Sub-rounded grains, black, brownish-black, brown, to brownish-yellow in color, are composed mainly of rutile, as identified by X-ray powder diffraction method. The black grains also contain a small percentage of ilmenite. The brownish-yellow grains are usually called leucoxene. Rutile is the only crystalline mineral in leucoxene as shown by its X-ray powder diffraction pattern. All these grains show pitted surfaces. The black grains have metallic luster, and are slightly magnetic. The size of the grains ranges from about 0.1 to 0.2 mm in diameter. Since no grains were found to possess the characteristic striae, prismatic form, and twinning of rutile, it is unlikely that rutile occurs as detrital particles in appreciable amount. All the sub-rounded grains may be regarded as leucoxene, an alteration product which is composed chiefly of amorphous titania and iron oxide, and some crystalline titania minerals (Allen, 1950).

As compared with a pattern of a fresh rutile sample from Magnet Cove, Arkansas, the X-ray powder diffraction lines of the rutile in the leucoxene of the brookite-bearing sandstone are generally broad and blurred and many weak lines are missing. This suggests that the individual grains of rutile are below 10^{-3} to 10^{-5} mm in diameter which is the optimum size for X-ray powder diffraction. It may also indicate that the sub-rounded grains of leucoxene contain a large portion of amorphous titania and iron oxide, instead of well-crystallized rutile. This amorphous titania and iron oxide represent the intermediate stage of the alteration of ilmenite to well crystallized rutile (Bailey, et al., 1956). Anatase or sphene was not identified in leucoxene.

Spessartite.—Most of the garnet grains are sub-angular though a few dodecahedrons were recognized. Based on an index of refraction of 1.81 the chemical formula of the spessartite is estimated to be $(Fe_{0.3}Mn_{0.7})_3Al_2Si_3O_{20}$.

Sphene.—A few yellowish grains which have high birefringence and strong dispersion were identified as sphene.

Tourmaline.—A few prismatic crystals of schorlite are present.

Zircon.—Zircon is the predominant detrital heavy mineral in the sandstone. Mauve and colorless varieties are most common, though pink and green varieties are present. Prismatic crystals are common but many grains do not show good crystal forms. The average size is about 0.1×0.16 mm. It has a dull brown fluorescent color under ultraviolet light. A Geiger counter test shows that the radioactivity of a concentrated zircon sample is about equivalent to 0.15 percent of U_3O_8.

MODAL AND NORMATIVE ANALYSIS

The chemical analysis of the brookite-bearing sandstone provides a basis for comparison of modal and normative analysis of the rock in which the unusually fine grain of some of the constituents prevents determination of relative proportion under the microscope. The data of the chemical analysis are listed in table 1.

Mode.—The modal composition of the sandstone is shown in table 2. Many of the mineral grains are hidden in the clay and are difficult to identify and differentiate un-

TABLE 1.—*Chemical analysis of the brookite-bearing Gallup sandstone of Cretaceous age. Sample No. 145-A-1*

Constituents	Weight percentage
SiO_2	27.61
TiO_2	35.36
Al_2O_3	9.48
Fe_2O_3	11.68
FeO	1.39
MnO	0.12
MgO	0.70
CaO	0.39
Na_2O	0.25
K_2O	0.40
P_2O_5	0.21
H_2O^+	5.65
H_2O^-	0.47
CO_2	0.00
ZrO_2	6.76
Total	100.47

Chemical analysis by Dr. H. B. Wiik, Helsinki, Finland.

der the petrographic microscope. Sphene is included with zircon. All the sub-rounded grains, whose color varies from black, brownish-black, brown, to brownish-yellow, are recorded as rutile. Thus, ilmenite, amorphous titania and iron oxide, and possibly some impurities such as clayey material, are included in rutile.

Norm.—The method of the calculation of the weight norm of the brookite-bearing sandstone is similar to that used for standard mineral molecules in the C.I.P.W. system of the chemical-mineralogical classification of igneous rocks (Johannsen, 1939, chapter 8). The method of the calculation of the molecular norm is similar to that discussed by Barth (1955). Nevertheless, the suite of normative minerals of the brookite-bearing sandstone is different from that of igneous rocks in that most of the normative minerals of igneous rocks are hypothetical mineral molecules whereas the normative minerals of the sandstone are actual mineral constituents which have been identified by various methods. However, ideal chemical formulae of the normative minerals are chosen for the calculation which provides a better estimate of the actual minerals than is revealed by the modal analysis (table 2).

TABLE 2.—*Mode of the brookite-bearing sandstone*

Mineral	Percentage
Zircon	13.50
Rutile	5.30
Quartz	8.20
Brookite, clay, etc.	73.00
Total	100.00

The normative minerals of the brookite-bearing sandstone are listed in the order of calculation in table 3. For instance, zircon is calculated first, and water is the last constituent to be calculated. The reader's attention is directed especially to the difference in the percentage of the molecular and weight norms. The percentages of the heavy normative minerals of the molecular norm are usually low, whereas corresponding percentages are high in the weight norm. For the light minerals the order is reversed. As the percentages of the molecular norm correspond closely to the percentages by volume, the percentage of brookite is calculated to be about 27.81 by volume and 33.57 by weight.

Magnetite, which occurs as a few individual grains and as molecules in ilmenite, is not calculated in the norm. Consequently the FeO of magnetite is included in ilmenite, and the Fe_2O_3 is included in limonite.

AUTHIGENESIS OF BROOKITE

Brookite is considered as an authigenic mineral in the Gallup Sandstone on the basis of the following criteria. First, it occurs chiefly as small boxworks between other detrital grains, and to a lesser extent in clay, mixed with rutile grains, or attached to some altered ilmenite. Second, in the absence of evidence of hydrothermal activity in the sandstone, it is believed that brookite was not introduced into the sandstone subsequent to deposition. Third, it appears that brookite was formed from the amorphous titania or the cryptocrystalline rutile of the leucoxene which in turn was formed at the expense of ilmenite. Sphene, which is not common in the sandstone, does not appear to have been the source of the titania for the formation of both rutile and brookite.

It has been pointed out that the sub-rounded, black, metallic, slightly magnetic grains contain rutile and a small amount of

TABLE 3.—*Norms of the brookite-bearing sandstone*

Normative minerals and their formulae	Molecular norm	Weight norm
Zircon $ZrO_2 \cdot SiO_2$	7.22%	10.00%
Apatite 3.33 $CaO \cdot P_2O_5$	0.43	0.49
Sphene $CaO \cdot TiO_2 \cdot SiO_2$	0.39	0.39
Albite $Na_2O \cdot Al_2O_3 \cdot 6SiO_2$	2.16	2.14
Orthoclase $K_2O \cdot Al_2O_3 \cdot 6SiO_2$	2.24	2.35
Montmorillonite* $(Si_8)(Al_{2.44}Mg_2)O_{20}(OH)_4$	8.67	6.52
Spessartite $(Fe_{0.3}Mn_{0.7})_3Al_2Si_3O_{20}$	0.37	0.39
Kaolinite $Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$	24.20	18.88
Ilmenite $FeO \cdot TiO_2$	2.44	2.79
Limonite $Fe_2O_3 \cdot nH_2O$	13.52	14.00
Brookite TiO_2	27.81	33.57
Quartz SiO_2	8.83	8.01
Water H_2O^-	1.72	0.47
Total	100.00	100.00

* The formula for montmorillonite having the highest possible Si:Al ratio (Brindley, 1951) is used as the normative formula. As silica only occurs in the tetrahedral layers and as all possible magnesium occurs in the octahedral layers, all the magnesia of the analysis is allotted to montmorillonite.

ilmenite, and possibly amorphous oxides of titanium and iron. The sub-rounded, brownish-black, brown, and brownish-yellow grains are composed largely of rutile, without any trace of ilmenite, and some amorphous material. These sub-rounded grains of various colors are believed to represent various stages of the alteration of ilmenite to leucoxene. All the grains show pitted surfaces which seem to be solution features. Minute yellowish-brown brookite plates were observed in some of the leucoxene under high magnification binocular microscope. Conclusive evidence is lacking for detrital rutile in the sandstone, although rutile has been reported as a detrital heavy mineral in Cretaceous sediments elsewhere in the San Juan Basin.

The rutile in leucoxene of the Gallup Sandstone is cryptocrystalline. Though it is possible that alteration of ilmenite to leucoxene (rutile) may have started prior to the deposition of the sandstone, it is believed that some of the change took place after the sediment was deposited. An authigenic mineral is sometimes defined as one having a crystallographic unit (Pettijohn, 1949, p. 497), formed *in situ* in sediments, and by sedimentary process. It is considered that the individual rutile in leucoxene of the Gallup Sandstone does not possess such crystallographic unit, because it is cryptocrystalline. Therefore, this rutile is not considered as an authigenic mineral.

When the sandstone is disintegrated, washed, and sieved into different sizes, it is found that individual boxworks and plates of brookite prevail below 100 mesh (0.147 mm in diameter). The boxworks tend to break down into plates during the stirring of the sample in a beaker. Thus, it is clearly shown that the brookite is not detrital and was formed subsequent to deposition of the sandstone. Because of the occurrence of the brookite plates in leucoxene, it is believed that the authigenic brookite is formed from the amorphous titania in leucoxene. It is difficult to determine whether or not the authigenic brookite could be altered from the cryptocrystalline rutile of the leucoxene. However, there is no doubt that the authigenic brookite was formed indirectly at the expense of the detrital ilmenite. Rutile-bearing leucoxene is common in sediments. However, the occurrence of brookite-bearing leucoxene and an abundance of authigenic brookite in the Gallup Sandstone of Cretaceous age is rather significant.

ACKNOWLEDGEMENT

The writers are grateful to Dr. Eugene Callaghan of the New Mexico Bureau of Mines and Mineral Resources who critically read the manuscript.

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FIGS. 1–11, SEPTEMBER, 1957

TRACING TURBIDITY CURRENT DEPOSITS DOWN THE SLOPE OF AN ALPINE BASIN[1]

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ABSTRACT

Black pelagic limestones were deposited in the sedimentation basin of the Nappe de Morcles (High Calcareous Alps) during the Upper Jurassic. This deep-water sequence, which generally reaches 100–150 meters in thickness, was interrupted at least nine times by coarse-grained clastic beds laid down by turbidity currents. The main constituents of these clastic beds are broken or intact reef-organisms, oolites, and clastic minerals, displaying well-characterized graded-bedding. This foreign material originated in the northern margin of the basin along which fringing reefs were growing. As these clastic beds are traced out to sea, as indicated by the regional paleogeography, they disappear one by one; moreover, peculiar topographic conditions allow the study of the textural modifications of a given bed at least 40 kilometers down the basin slope, along the inferred main direction of flow of the turbidity currents. Particularly interesting is the behavior of the apparent maximum diameter of a given clastic component (index of clasticity) measured at the bottom of the bed, since it may reveal the horizontal grading along the direction of flow. The variations of clasticity along the inferred direction of flow of turbidity currents are in general agreement with the experimental data as well as with the mechanics admitted for this peculiar type of sedimentation. However, further studies are required to ascertain their possible use in correlation.

INTRODUCTION

Since the importance of turbidity current deposition has been recognized in geosynclinal clastic sedimentation, especially in graywackes, the problem of correlating these peculiar beds appears of high interest to petroleum geology.

Inasmuch as these beds wedge down the basin slope because of decreasing velocity and increasing dilution of the turbidity currents, we are interested in the textural modifications along the inferred main axis of flow of the currents, in what we may call "longitudinal correlations." Transversal ones are more difficult to handle, in a general way, because of turbulent behavior of the marginal parts and also because they depend largely on the distance separating starting points along the shoreline.

In such an approach, it is necessary to try to use any variable, and if possible independent variables in order to get to a clear picture of the problem. In the Upper Jurassic of the High Calcareous Alps of France, I (Carozzi, 1955) have studied an example which is perhaps a peculiar one because the turbidity current sediments are calcarenites. Moreover their sedimentary variability and the size of the currents are not so large as in graywacke sedimentation and thus it is possible to get a general picture in a relatively small area.

At first glance, it may appear strange to find turbidity currents in the alpine foredeep or helvetic trough where usually marly limestones and shales were deposited under conditions never corresponding to excessive sedimentation. But the Upper Jurassic was a period of tectonic instability in the Western Alps and such a situation generated sub-marine slides which evolved down the basin slope in widespread turbidity currents.

Let us now outline the general frame of the problem (Collet, 1943). The sediments of the basin we are studying build up today the Morcles Nappe which is the lowest tectonic element of the High Calcareous Alps (fig. 1). During Upper Jurassic times, the basin reached a width of 60 km and was bordered both to the northwest and the southeast by two old crystalline massifs: the Aiguilles Rouges and the Mont Blanc massifs. They are formed by granitic batholiths surrounded by crystalline schists in which they were intruded during the Carbo-

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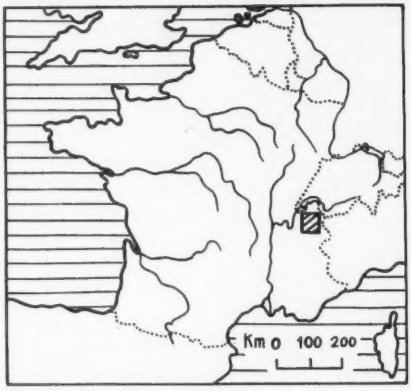

FIG. 1.—Map of France showing location of the area studied.

niferous. Intense peneplenation occurred during Permian times and, from then on these massifs were incorporated as rigid blocks in the alpine orogenic belt. Shallow waters encroached them from time to time, however, and in the Upper Jurassic the uplift was strong enough to raise them out of marine conditions and they were covered by brackish and fresh-water lakes separated from the open sea by a large but discontinuous zone of fringing coral-reefs.

During the main orogenic alpine phase, in Middle Oligocene, the Mont Blanc was thrusted toward the northwest over the basin of sedimentation, the filling of which was pushed in the same direction, as a nappe, overriding the Aiguilles Rouges block.

The basin is today reduced to a root-zone of 2 kms width whereas the Morcles Nappe thus formed shows an overthrust of 20 kms.

Since most of the turbidity currents originated from the northwest coast-line of the basin, that is from the southern slope of the Aiguilles Rouges massif, the sections located near the old coastline are to be found today in the tectonic zone called "the Autochthonous," but the sections situated down the basin slope are included today as overturned sequences in the arch-bend and the reversed limb of the nappe (fig. 2).

METHOD OF INTERPRETATION

The essential characteristic of the author's method (Carozzi, 1950) is to express by envelope-curves, drawn parallel to the stratigraphic column, the variations of the greatest possible number of mineral and organic micro-components as well as the results of chemical analysis, in order to give an objective and absolute description of the facies and the relations to their environment (depth, distance to shore line, currents.)

The study of the detrital minerals or of any inorganic or organic fragment showing a similar behavior is first done by the measurement of the variations of the maximum diameter of the grains. The latter is a direct function of the power of transporting agents, being thus a trustful indicator of the distribution and intensity of sub-aqueous currents.

Of course, in thin section, only the apparent maximum diameter of a given mineral (usually quartz) can be obtained, but the difference between this apparent diameter and the true maximum one (shown by mechanical analysis) is slight when one takes the indispensable caution of computing a

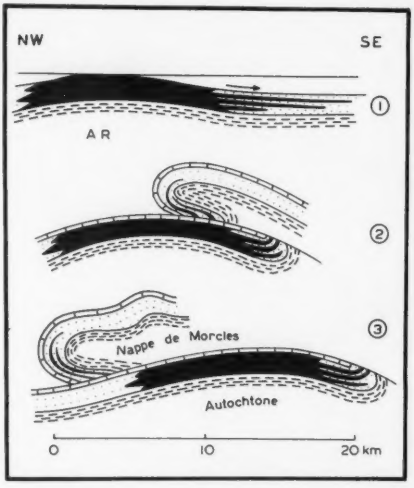

FIG. 2.—Diagram showing the tectonic effects of overthrusting on the deposits of turbidity currents in the Upper Jurassic of the Haute-Savoie. 1. Emplacement of clastic beds derived from the coral reefs of the Aiguilles Rouges (A.R.). Only three beds are shown. Reefs and clastic beds in black, pelagic deposits dotted. 2. Initial stage in the development of the Nappe de Morcles. This must have started after emplacement of the clastic beds. 3. Present position of the Nappe de Morcles after a displacement of about 20 km over the Aiguilles-Rogues and its autochthonous cover.

minimum of one hundred grains. The apparent maximum diameter corresponds to the so-called "clasticity index" of the sample for the given mineral. The classification of the facies according to their median maximum clasticity expresses the degree of agitation of the water or the intensity of subaqueous currents, but the depth criterion can be given only by micropaleontological data.

The second way to study detrital minerals is to measure their frequency, that is, the absolute number of grains of a given mineral in a given volume of the rock. This measurement is independent of clasticity measurements. Practically, the grains are computed on a carefully measured surface of the thin-section; in the present study, a square surface of 18.2 mm side. This surface can be enlarged or reduced for special purposes provided the convention is clearly established.

When authigenic minerals appear as crystals or grains, measurements of clasticity and frequency can be applied to them. However, their significance is entirely different; the clasticity loses its sense and is only the expression of the crystallographic growth of the grains, whereas their frequency reveals the intensity of the process of formation. Pyrite pigments are computed by establishing a set of thin-sections corresponding to different amounts of iron, chemically controlled, building up a sequence for instance from 0 to 10 degrees of iron content. The value corresponding to a given thin-section may be established by optical comparison. In this fairly uniform sequence such a precision is largely sufficient for routine work.

Microfaunas are studied usually by frequency measurements, sometimes by size, for checking environment changes or transportation. The behavior of the benthonic microfaunas affords the bathymetric criteria which are to be used for the interpretation of the curves of the minerals in order to reach the final picture of the environment.

In the present study, the following curves are illustrated:

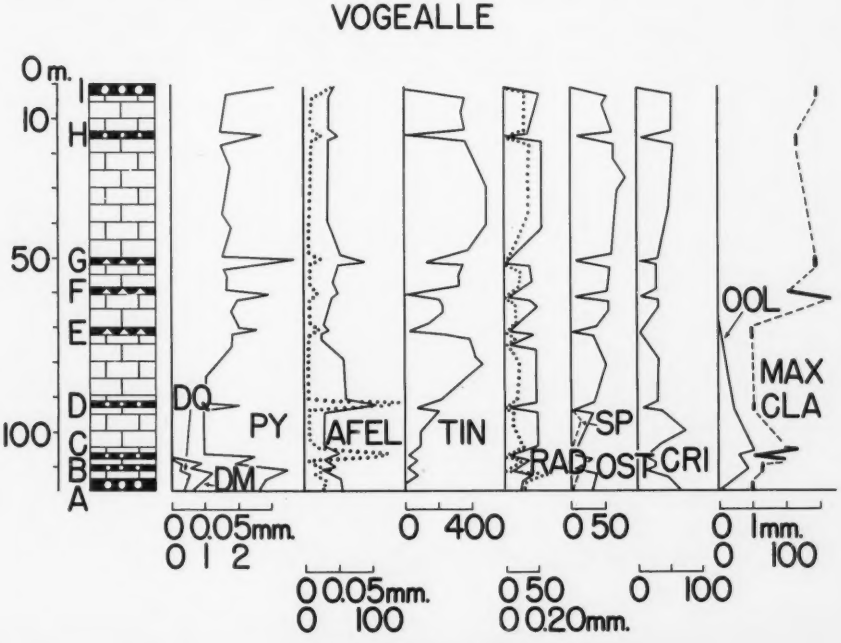

FIG. 3.—Section of Vogealle.

clasticity of quartz grains (DQ)
clasticity of muscovite flakes (DM)
frequency of pyrite pigments (PY)
diameter of authigenic felspars (AFEL), solid line
frequency of authigenic felspars (AFEL), dotted line
frequency of Tintinnoids (TIN)
diameter of Radiolaria (RAD), solid line
frequency of Radiolaria (RAD), dotted line
frequency of Ostracods (OST)
frequency of pelagic Crinoids: Saccocoma (CRI)
frequency of Sponges (SP)
frequency of reworked oolites (OOL)
clasticity of inorganic grains in the calcarenite beds (MAX CLA) expressing the internal texture.

The upper limit of all the columns corresponds to the upper limit of the Upper Jurassic, the thicknesses are measured from this line downward; triangles and circles on a black background symbolize microbreccias and microconglomerates respectively.

PETROGRAPHIC DESCRIPTION

The Upper Jurassic is made up of a thick sequence (100 to 150 m) of black, fine-grained pelagic limestone showing by chemical analysis between 97 and 98% calcium carbonate. Its fauna is entirely pelagic, made up of Ammonites, Tintinnoids, Radiolaria and floating Crinoids (Saccocoma). Clastic minerals are rare and usually found at the bottom of the sections; they are quartz and muscovite.

On the other hand, authigenic minerals, calcedonite and felspars, are locally well developed as well as pyrite concretions. Statistical measurements of both organic and inorganic components show that the conditions of sedimentation were almost uniform all over the Upper Jurassic; some evolution is shown by the behavior of the Radiolaria and Tintinnoids but cannot be generalized.

This uniformity of conditions corresponding to what we may call the "fundamental or autochthonous facies" was interrupted several times by clastic beds, the thickness of which ranges from some millimeters to some decimeters, which are abruptly interstratified in the fine-grained limestones and which are directed by completely independent laws (allochthonous facies).

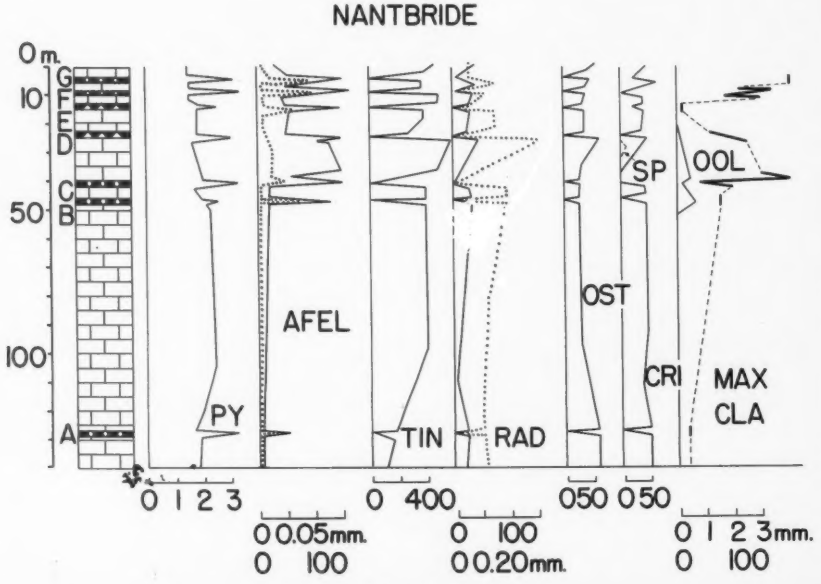

FIG. 4.—Section of Nantbride.

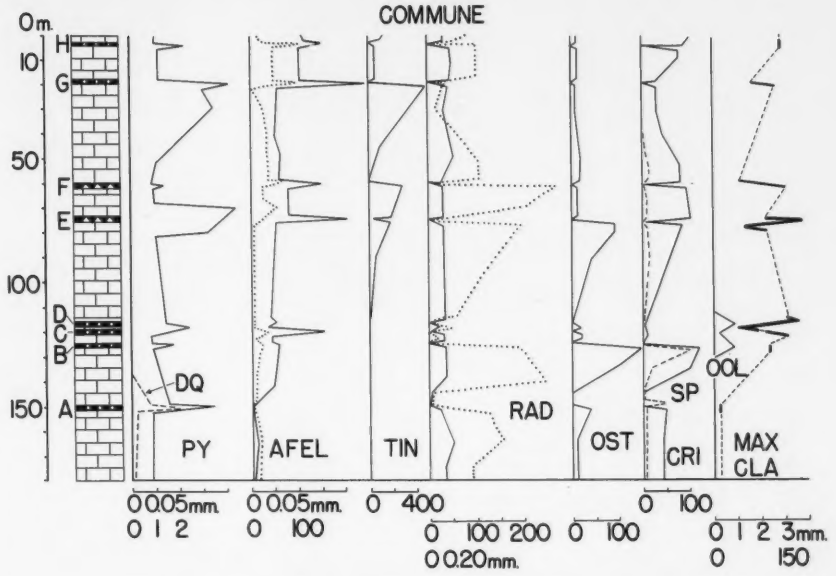

FIG. 5.—Section of Commune.

Petrographically, these beds are microbreccias and microconglomerates made up by different types of elements, both inorganic and organic. The first group includes entire or broken oolites, well-rounded pebbles of oolitic limestone, and relatively angular fragments of the fine-grained fundamental facies which seems to have been picked up from the neighboring regions.

Clastic quartz grains, clay, iron oxides and sulfides, and authigenic felspars associated with calcite build up the matrix. Organic fragments are rounded or angular according to the considered bed; they build up a homogeneous faunal assemblage characteristic of reef as well as brackish and fresh-water conditions. None of its components is to be found in the fundamental facies, they originate from a completely different environment. This faunal assemblage includes fragments of: Rudists, Brachiopods, Crinoids, Corals, Bryozoans, Dasycladaceae and Characeae, Annelids, and numerous arenaceous Foraminifera living under reef conditions.

Different textures are shown in the clastic beds, some of them reveal a chaotic or uniform texture, others show just an incipient sorting whereas the bulk of them have a well-developed simple or composite graded-bedding. The latter is made up of several layers, each showing graded-bedding, and an irregular erosional lower limit. Such a composite graded-bedding is considered to originate through the interference of two or more turbidity currents. For instance, the front of the second flow advancing more rapidly than the first, caught its tail and plunged underneath it. Thus, a second cycle of deposition began before the first got to an end and was preceded by erosional processes.

The distribution of the organic fragments in the clastic beds is highly influenced by the graded-bedding. That is, the coarse zones at the lower parts of graded beds contain fragments of Rudists, Brachiopods, Corals and Bryozoans associated with large Foraminifera and isolated oolites. This is a pseudo-fauna made up only of the components which were originally of large dimensions (Foram's) or which could be broken only along definite partings (Bryozoans tubular cells, Rudists prisms . . .). On the other hand, the fine-grained zones of the upper portion of the graded-beds reveal

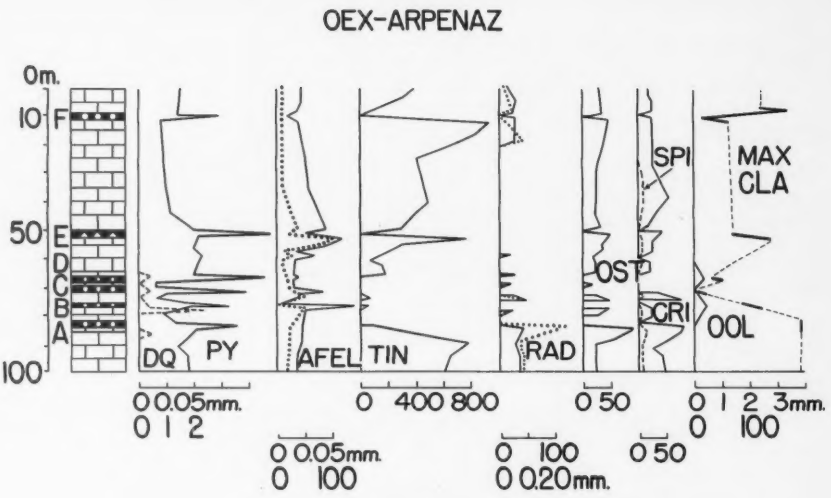

FIG. 6.—Section of Oex-Arpenaz.

fragments of Echinoids, Crinoids, Annelids, small Foraminifera and green algae, originally of smaller dimensions. However, fragments of the fine-grained pelagic limestone, the fundamental facies, are distributed in about the same proportion in all the beds because their dimensions are not influenced by any internal structure; they can thus give rise to pebbles of any possible dimension.

The lower contact of the clastic beds is usually irregular but clean-cut and shows locally typical load, flow and drag casts. When the clastic beds have a chaotic or uniform texture, their upper limits are as clean-cut as are their lower, the matrix is slightly developed, and no contamination whatsoever can be observed, as if the deposition stopped as abruptly as it started, the clastic bed being rapidly covered by the fundamental facies.

When an incipient sorting is provided, the proportion of the matrix increases; pelagic Radiolaria and Tintinnoids can be observed scattered in the lower parts of the beds as well as fragments of the fundamental facies with pull-a-part structures indicating mechanical contamination during transportation.

When graded-bedding is well developed, the proportion of the matrix increases upward and gives rise to a gradual transition toward the fundamental facies; pelagic microfauna is abundant all over the matrix but especially in the upper part.

STRATIGRAPHIC COLUMNS

The studied sections from the coastline down the inferred slope are:

1.—*The Vogealle* section (fig. 3), nearest the shore line, probably 1 km, is 115 m thick and shows 9 clastic beds; those at the top and bottom of the section are microconglomeratic, those in the middle are microbrecciated. Most of them are not graded; one is graded and another has composite graded-bedding. Oolites are more abundant in the lower beds than anywhere else.

Clastic quartz and muscovite are limited to the lowest part of the section. Pyrite and authigenic felspars frequencies show a relative uniform curve, expressing the local pelagic facies, abruptly cut 9 times by high peaks corresponding to increased amount of iron and silicate solutions or colloids brought by the turbidity currents at the same time as the clastic elements and which gave rise to neo-formations. All the frequency curves of pelagic organisms, Tintinnoids, Radiolaria, Ostracods, and Crinoids, show important minima or even total absence when clastic beds were deposited.

They nevertheless show a general evolution corresponding to an increase of frequency upward.

2.—*The Nantbride* section (fig. 4), 4 km down the slope, 3 km from the preceding section, has a thickness of 140 m and shows 7 clastic beds largely concentrated in its upper portion. However, microbreccias are still preceded and followed by microconglomerates. Nearly half of the beds show simple or composite graded-bedding which represents a much higher proportion than in the preceding section. Oolites seem to be concentrated in the lower part of the upper portion. Frequency curves of pyrite and authigenic felspars increase upward as do the pelagic Tintinnoids. The other components have a uniform behavior.

3.—*The Commune* section (fig. 5), 6 km down the slope, 2 km from the preceding section, has 8 clastic beds (one in double) more or less regularly distributed. Here again microbreccias are preceded and followed by microconglomerates. Five beds out of eight have graded-bedding normal or composite, the oolites are here also concentrated only in certain beds. Stability is almost generalized for the other components.

4.—*The Oex-Arpenaz* section (fig. 6), 21 km down the slope, 15 km from the preceding one, has 100 m thickness and is interrupted by 6 clastic beds which are thinner and less characteristic than they are in all the preceding sections; microconglomerates still surround microbreccias. All but one of the beds show very well-developed graded-bedding. Thus, in spite of the fact that 3 beds have wedged out from the coast to this point, the graded-bedding clearly increases. Oolites are limited to some beds only. Pyrite and authigenic felspars frequencies diminish upward but the other components are stable.

5.—*The Giettaz* section (fig. 7), 40 km down the slope, 19 km from the preceding section, reveals only two clastic beds in its 140 m thickness; one has an incipient grading and the other shows a composite graded

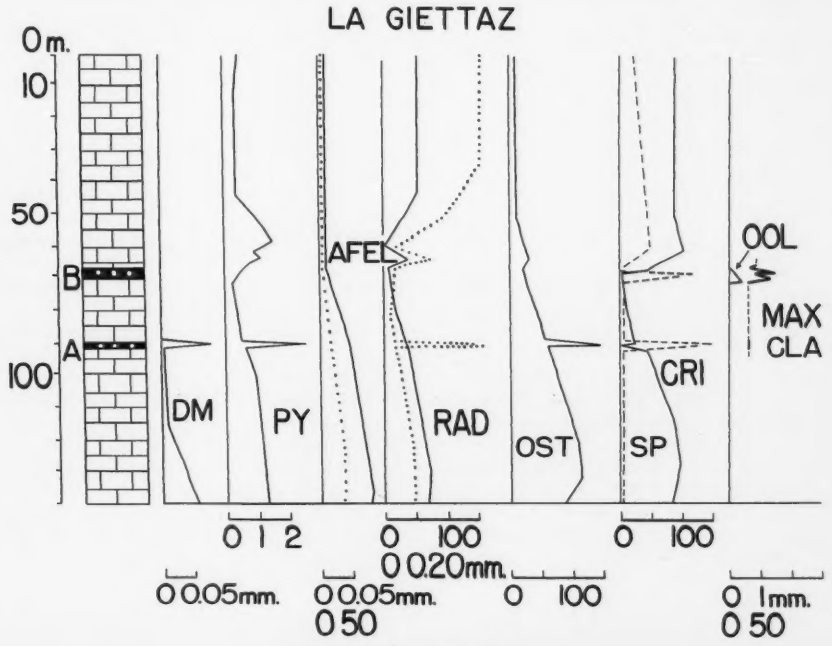

FIG. 7.—Section of La Giettaz.

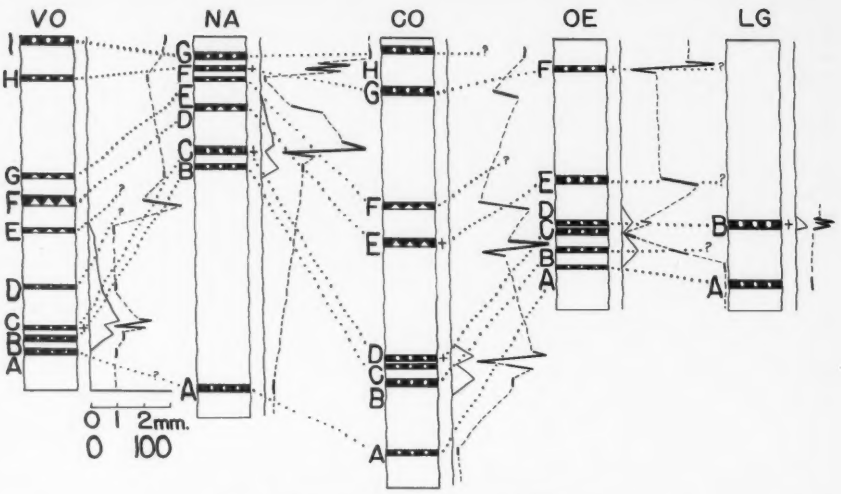

FIG. 8.—Tentative correlation of the clastic beds.

bedding and a high concentration of oolites. Mineral components tend to diminish upward as do the Tintinnoids among the organisms. This is the deepest section down the basin slope.

TENTATIVE CORRELATION

The correlation of the clastic beds across the five studied sections must be worked out from general correlations of groups of beds toward more precise correlations bed to bed when possible. The following criterions appeared useful:

1.—*The facies of the clastic beds:* microbreccias or microconglomerates which allow a first grouping of the beds.

2.—*Texture of the clastic beds:* lack of sorting, normal or composite graded-bedding lead to a further subdivision inside the groups already delimitated by facies. However, one must not forget that a given bed may start with a chaotic or uniform facies and evolve gradually down the slope in a well-graded one. It may also show composite graded-bedding if it happened to interfere with another turbidity current deposit.

3.—*Distribution of peculiar organic or inorganic components:* it happens that a given turbidity current or a group of them may contain organic fragments or mineral elements which are characteristic because they were probably afforded only in a limited zone of the coastal sediments or during a given amount of time. In the present study, oolite occurrences, as well as fragments of *Clypeina jurassica*, a marine green algae, were very useful.

Facies (microbreccias and microconglomerates) allow a primary subdivision of the clastic beds into 3 main groups. They are, from bottom to top, in the Vogealle section: 1 to 4 (A, B, C, D) 5 to 7 (E, F, G), 8 and 9 (H, I). Their horizontal equivalents in the other sections are shown in figures 8 and 9.

The beds of the first group (microconglomeratic) are the only ones containing reworked oolites; very often they show composite graded-bedding, indicating interferences between currents. Bed A (no. 1), uniform in texture or showing incipient grading in some places, can be traced easily in all the sections; it also lacks oolites. Bed B (no. 2) is uniform until Oex-Arpenaz where it is graded-bedded before wedging out. It shows a first maximum in oolites frequency curve. Bed C (no. 3) is present everywhere either with composite graded-bedding or made up of two graded-bedded and distinct layers showing complex interferences; it can be traced in all the sections and corresponds to the second and last peak of oolites frequency. Bed D (no. 4), the last of this lower group

is just shown in the Vogealle section and is of little importance. Similar conditions are found in the layer E (no. 5), lowest element of the medium microbrecciated group. Bed F (no. 6) shows a well-developed graded-bedding and can easily be traced in all the sections but the last; it is a layer displaying a remarkably constant character. Bed G (no. 7) has a uniform texture in two sections and shows graded-bedding just before wedging out in the Commune column.

The upper subdivision includes two microconglomeratic beds; bed H (no. 8) starts uniform then becomes graded-bedded in a normal or composite way, whereas the last bed I (no. 9) has a uniform texture in the three sections in which it can be traced. Important observations may already be pointed out:

1.–The thicknesses of the fundamental facies separating correlated clastic beds show sometimes important variations which must be attributed to sub-marine topography of the basin slope.

2.–The type of facies (microbreccias or microconglomerates) remains constant all over a given bed or a given set of beds.

3.–When a clastic bed is traced down the basin slope, it may keep a given texture or may start uniform and gradually evolve in normal graded-bedding.

4.–Composite graded-bedding in one thick layer, indicating interference of several currents, may be replaced down the slope by several distinct layers as if the different turbidity currents had resumed their individuality after a period of mixing. The reverse situation has not been met but may be possible theoretically.

HORIZONTAL GRADING

Horizontal grading takes place only if the mechanism by which the sediments are spread out over the bottom or slope of the basin has produced sorting; thus it must be looked for only in clastic beds showing graded-bedding both with microbrecciated or microconglomeratic facies.

When a turbidity current starts in the upper part of the basin slope, it is usually loaded to capacity or even overloaded. It

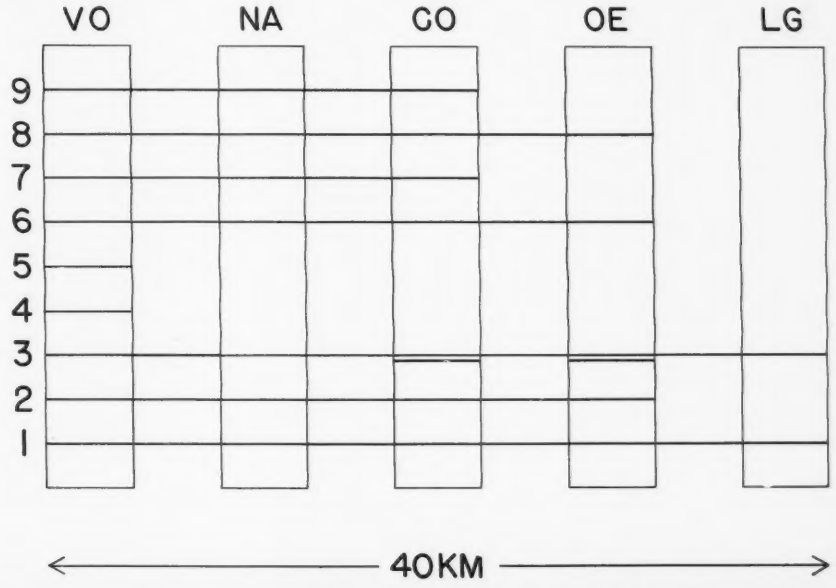

FIG. 9.—Schematic distribution of the investigated turbidity currents.

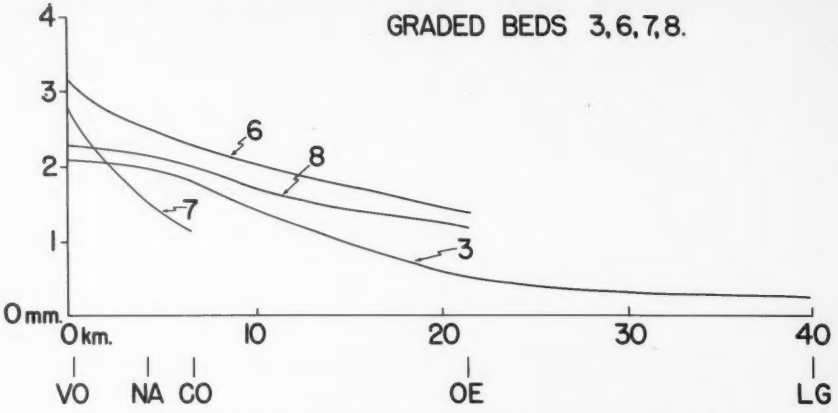

FIG. 10.—Behavior of clasticity (horizontal grading) in clastic beds deposited by single turbidity currents.

starts to deposit the larger grains and, as it continues down the slope, allows grades of decreasing size to settle farther and farther along. (Kuenen, 1951; Natland and Kuenen, 1951; Kuenen and Menard, 1952).

This decrease is partly due to loss of larger grains, leaving only smaller ones to be dropped, and partly to decreasing velocity and increasing dilution which allow fractions to settle which before could still be transported. Thus, the horizontal grading is produced. However, this process is a slow one and the lateral change of grain size of the bed is stretched out over great lengths. In nature, the horizontal grading tends to be imperceptible except in relatively coarse beds deposited by currents of moderate volume. These conditions seem to be met by the investigated examples.

Horizontal grading can be readily expressed by the measurement of the "clasticity index" taken at the bottom of the graded beds because the coarsest elements are indicators of the maximum strength reached by the turbidity current in a given spot.

Beds no. 3, 6, 7, and 8 give interesting curves (fig. 10) showing definite and gradual decrease of clasticity down the slope. Beds no. 3 and 6 display normal or composite graded-bedding over their entire length, whereas beds no. 7 and 8 start uniform at Vogealle but are graded-bedded in all the other sections.

With materials having high internal friction, sliding is predominant and wedges of sediments are laid without grading in a vertical or horizontal direction as shown by bed no. 9 (fig. 11). Many turbidity currents have originated in that way along the upper portion of the slope and acquired their characteristics only farther down. These types might be considered as normal inasmuch as only one current is involved or only a group of currents following the same trend and generated in short time interval. Abnormal types correspond to a situation in which two or more turbidity currents, coming from different spots along the shoreline, happened to interfere somewhere down the slope. In the case of graded beds, the first gradual decrease of clasticity down the slope is followed by a rapid increase (due to mixing) leading to a second and similar decrease downward. Bed no. 1 (fig. 11) is such an example showing that at 20 km from the coast a powerful turbidity current reached the end deposition zone of another one and took its place.

In the case of beds deposited by sliding and lacking sorting, the example is shown by bed no. 2 (fig. 11), two horizontal lines are connected by a curved line also generated through mixing.

It is interesting to note that in both cases of graded or uniform beds, the portion of the curve down the slope starts with values higher than those of the up-slope portion

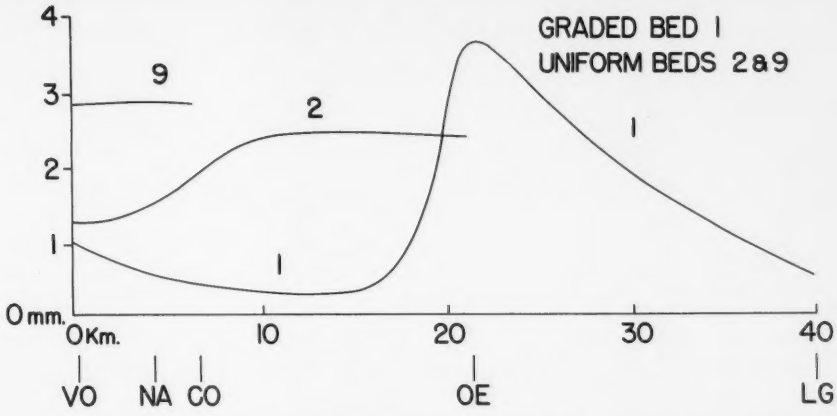

FIG. 11.—Behavior of clasticity in clastic beds deposited by sliding and through interference processes between turbidity currents.

thus indicating that the second current or slide was always larger and more powerful than the first one.

Sanders (1956) has worked out a schematic diagram, based partially on an actual example, corresponding to a theoretical slope of a basin along which turbidity currents are moving. In the littoral zone, about 10 km wide, no grading is apparent in the coarse conglomerates and sandstones which are deposited mainly by sliding phenomena. Down the slope, the proximal facies gradually merges into medium to fine-grained sandstones in which graded-bedding is characteristically displayed for about 30 km. Then, at the distal end of the turbidity currents, that is at the bottom of the basin slope, grading disappears and the deposits are very fine-grained sandstones or lutites.

Our observations are easily tied in such a diagram, the proximal zone with slumping features has not been observed, but the curves (figs. 10, 11) illustrate the main portion of the basin slope and its bottom where the turbidity currents disappear.

CONCLUSIONS

Much still has to be worked out before interference processes between turbidity currents will be entirely cleared, however, we think that the normal horizontal grading in turbidity current deposits is met more frequently than theoretical studies tend to ascertain, and that it represents an interesting feature for correlations and for unravelling the direction of displacement of currents and slides.

ACKNOWLEDGMENTS

The writer wishes to thank his colleagues Professors J. L. Hough, J. E. Sanders, and H. R. Wanless for their constructive criticisms.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 27, NO. 3, PP. 282–292
FIGS. 1–12, SEPTEMBER, 1957

OBSERVATIONS ON LIMESTONES[1]

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ABSTRACT

This paper discusses the origin of pyrite and ankeritic dolomite in certain Pennsylvanian limestones from Iowa, as well as evidence for the clastic nature of the limestones. It is concluded that:

1.—Pyrite can form in bottom sediment within a few centimeters of the depositional interface in spite of a flourishing benthonic fauna;
2.—Ankeritic dolomite may form in the same environment as, and contemporaneous with, pyrite;
3.—All the limestones studied show evidence of clastic origin, including abrasion of the fine-grained calcite matrix.

INTRODUCTION

During the fall of 1954 the writer had occasion to sample the limestone units of the Des Moinesian Series which crop out in Dallas, Guthrie, and Madison counties, Iowa. Subsequently a study of the material was made in order that some of the physical characteristics of the limestones might be set forth, and also to determine the extent to which algae and algal activity have been responsible for their accumulation. As the work progressed two features of the limestones became conspicuous: (1) their content of pyrite and ankeritic dolomite; and (2) their clastic nature.

The purpose of this paper is to discuss limestone genesis in terms of these features and to set forth the writer's interpretation of environment as suggested by them.

PYRITE

The presence of pyrite in almost every limestone sample studied would at first appear to be somewhat peculiar in view of the fact that the limestones are marine, carry a marine fauna, and are generally believed to be the product of sedimentation in an open-circulation environment.

Pyrite occurs in these limestones as minute euhedra or as clusters of euhedra, randomly distributed (figs. 1 and 2). Close observation reveals that in many cases the mineral is more abundant along the boundaries of fossil shells (figs. 3 and 4), and in a few cases it is concentrated along the ceilings of whole brachiopod shells (fig. 5) (since the stratigraphic orientation of the samples was known, it was possible to distinguish the original position of entombment of fossil remains). The euhedra predominantly vary between .005 and .01 mm in diameter, but are often of such dimensions as to be barely discernible at 400 magnifications. When abundant the pyrite may preferentially replace foraminiferal debris or fragments of bryozoa (figs. 6 and 7). In the only positive instance of preservation of plant remains it was noted that these remains were largely pyritized and also that the rock contained a super-abundance of pyrite.

The occurrence of syngenetic pyrite has been noted by many authors. Krumbein and Garrels (1952) point out that pyrite and calcite can form and be stable in an environment in which the pH is approximately 8.0 and in which the Eh is approximately −0.3. It is important to note that the negative Eh value does not imply a stagnant environment. Marine open-circulation conditions may exist down to the depositional interface, while below this level there may be a tendency toward a reducing environment due to depletion of oxygen. They state the problem thus (Krumbein and Garrels, 1952, pp. 31):

> "... the mud-bottom beneath neritic normal marine open-circulation environments may have the property of the euxinic environment."

However, the *actual capacity* for reduction will depend on the presence of organic matter and the *actual capacity* for formation of syngenetic pyrite will be dependent on the concentration of S ion and iron.

If we are dealing with an open-circulation environment it is very unlikely that negative Eh values necessary for the formation of pyrite would exist in the bottom waters above the depositional interface. It cannot be assumed that the bottom waters were so

[1] Manuscript received August 20, 1956.

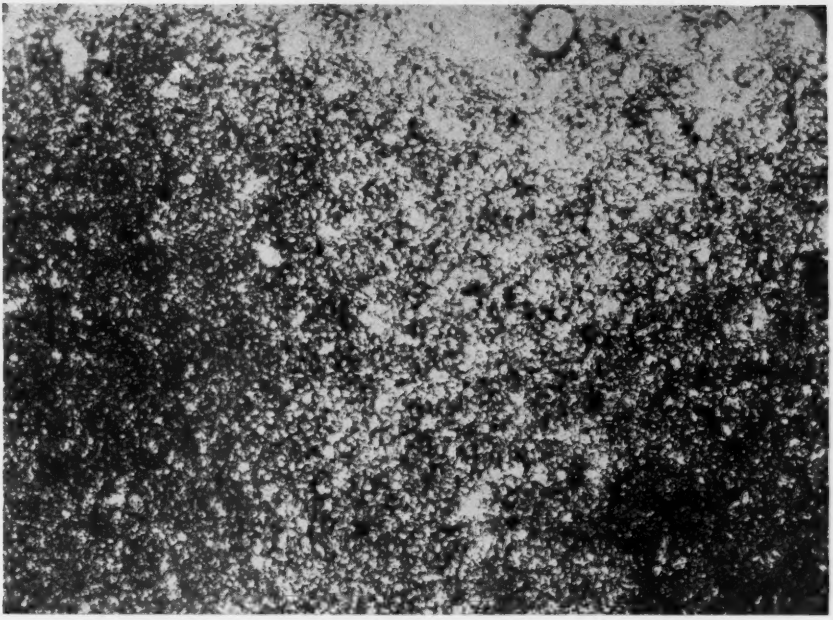

FIG. 1.—Ardmore Limestone. R. 29 W., T. 78 N., SW ¼ sec. 9. (Union tp., Dallas Co.) Granular limestone, showing disseminated grains of pyrite. Most of the matrix is granular calcite, but some fragments of shell material occur. Photomicrograph ×80.

poorly aerated as to be reducing in nature, for such a condition would be restrictive and inhibitive to the fauna represented in the limestone, especially foraminifera. Therefore, we must turn to a post-depositional or penecontemporaneous origin for the pyrite.

The clastic nature of all the limestones studied is indicative of near shore accumulation and implies a rate of deposition sufficiently rapid to entomb at least a small amount of organic matter. Actually, in any basin of deposition fringed with vegetation and supporting a substantial fauna, it is to be expected that varying amounts of organic matter are incorporated in the bottom sediment, and persist for some time after burial, even if sedimentation proceeds at a very slow pace. Tyler[2] has observed a black sludge forming on the undersides of small boulders near the shore of a fresh water lake, where no sediment was being accumulated. This sludge was not analysed, but since it was observed to rapidly oxidize to a brown color on exposure to the atmosphere, it was concluded by him to be an iron sulphide.

If we assume the depositional interface and the zero Eh level to be coincident, i.e. above the depositional interface oxidizing conditions exist, while below the depositional interface reducing conditions exist, then the decomposition of the entombed organic matter would be anaerobic. Such decomposition would yield a variety of products including H_2S, thereby imparting reducing capacity to the environment and providing S ion. These conditions having been satisfied, pyrite could only have formed if sufficient iron were being introduced to the sea at the time of accumulation of the limestone. Near-shore accumulation indicated by the clastic nature of the limestones implies reasonable proximity to mouths of river systems so that if iron were being in-

[2] Dr. S. A. Tyler, Department of Geology, University of Wisconsin; Personal communication.

troduced to the sea, the areas of carbonate deposition would be within the limits of readily available iron. This reasoning is in agreement with the beliefs of many writers that iron in any form does not occur as a prominent accessory constituent of sediments accumulated far from shore except under very unusual conditions.

The requirements necessary for the formation of pyrite can exist, then, within a few centimeters of the depositional interface, constituting no impediment even to the sessile and vagrant benthonic fauna. The actual state in which the iron existed at the time of its introduction into this environment is not important; it could only persist in the reduced state, and to this end trivalent iron would subsequently accommodate itself. The frequent occurrence of euhedral pyrite along shell boundaries as well as randomly dispersed throughout the body of the rock suggests that the iron was in solution or in colloidal form in the interstitial fluids of the unconsolidated bottom sediment, and that the pyrite formed under very marginal conditions. Indeed, satisfaction of the postulated conditions seems to have been so local in many cases that whole brachiopod shells have a concentration of pyrite crystals along their ceilings where the H_2S evolved from decaying animal matter within the shell was trapped.

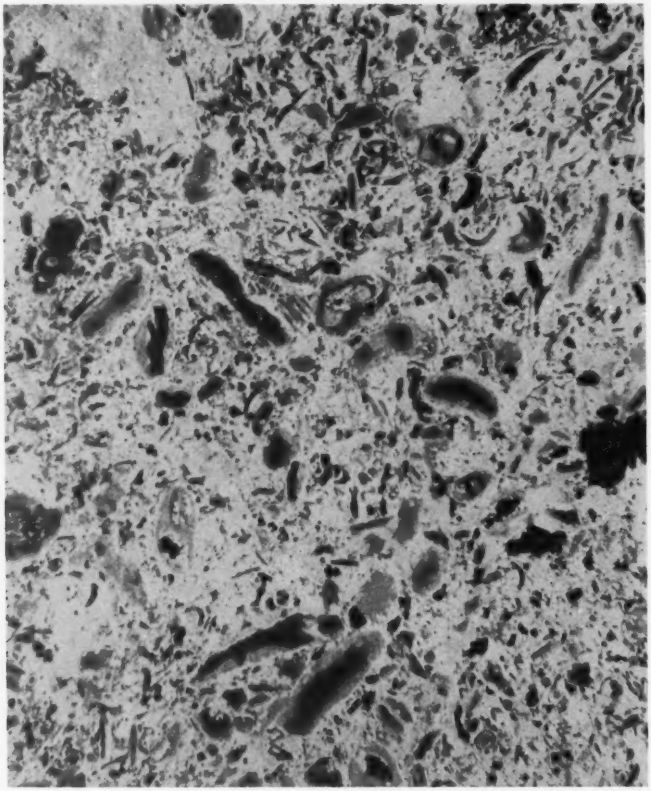

FIG. 2.—Coal City Limestone. R. 30 W., T. 79 N., SW ¼ sec. 25. (Jackson tp., Guthrie Co.). Calcarenite; fragments of all sizes cemented together with finely granular calcite; ankeritic dolomite and pyrite are common in the interstices. Some smaller fragments bear abraded algal encrustations. Negative print ×8.

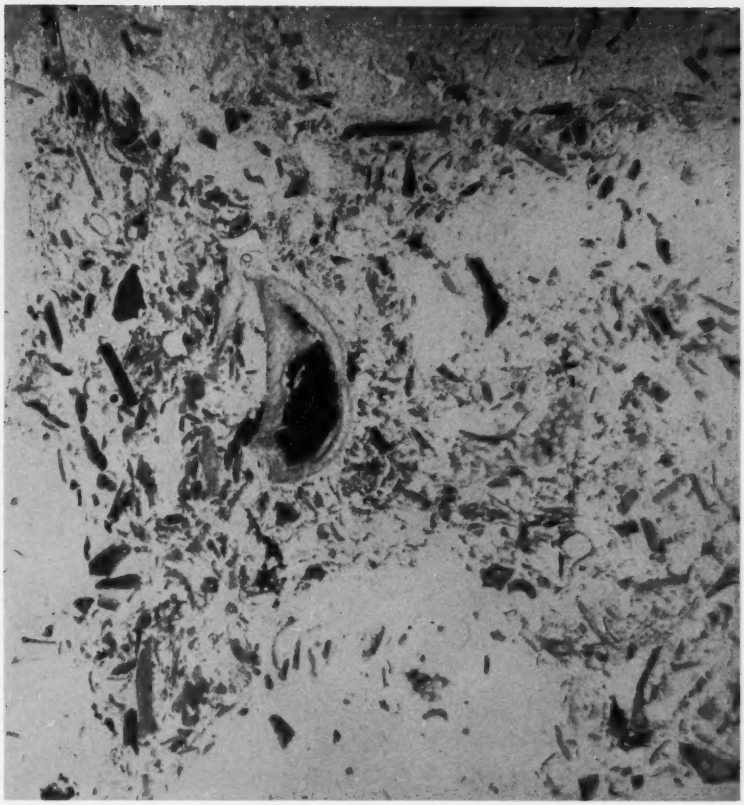

FIG. 3.—Myrick Station Limestone. R. 26 W., T. 75 N. (South tp., Madison Co.) Highly angular unsorted debris in an argillaceous, finely granular limestone. Pyrite is abundant, but difficult to distinguish in this photo; the change in color toward the top is due to a decrease in clay content. Negative print ×8.

Rubey (1930) has described a pyritiferous limestone from the Greenhorn Formation of Wyoming, in which pyrite, gypsum, pyrrhotite, and bone phosphate are associated. The pyrite occurs in minute nodules, finely crystalline, or as coatings on shell fragments. He appeals to waters escaping from underlying black shales during compaction. From a study of the rates of sedimentation, he concludes that the most satisfactory conditions for finding pyrite and limestone would be those of relatively shallow water, a rapid rate of sedimentation, and the burial of organic matter, thereby allowing formation of sulphides in the putrifying ooze, chiefly below its upper surface. The writer agrees with the conditions postulated with these modifications:

(1) is is not necessary that the bottom sediment be "putrifying" in the strict sense of the word, for sulphides can form in the presence of very little organic matter; and

(2) if proper conditions exist within the calcareous mud, then sulphides can be expected to form without outside stimulation or inducement in the form of fluids from underlying black shales. It is interesting to note that Rubey feels that lime-secreting bottom-living organisms might grow and be preserved at or above the interface while

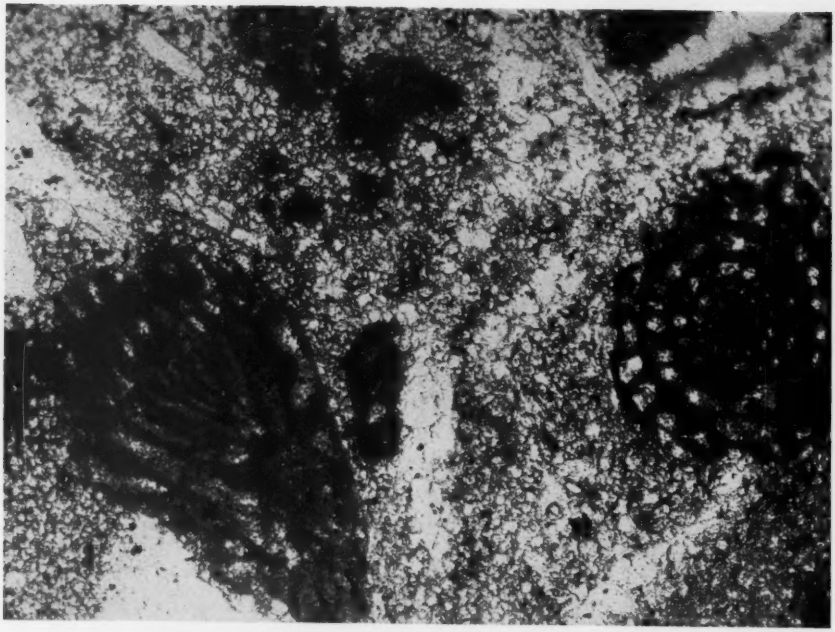

FIG. 4

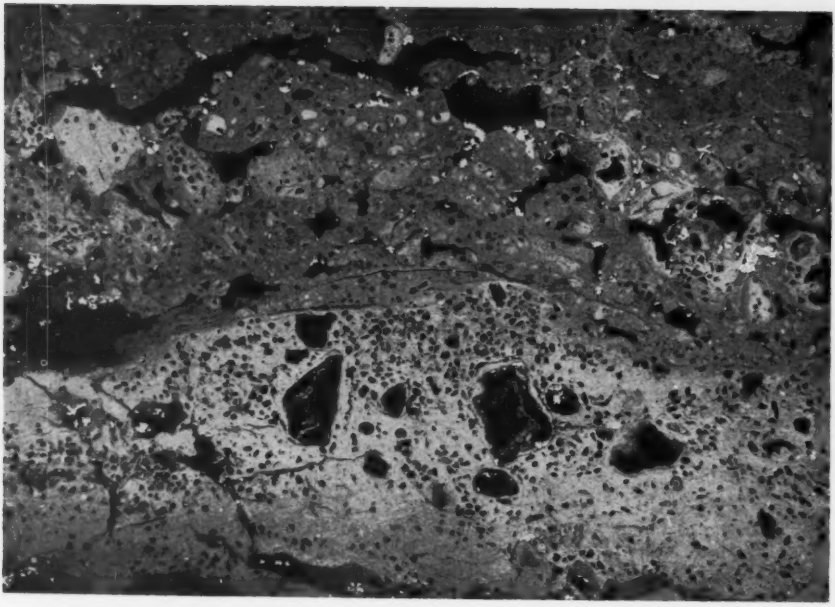

FIG. 5

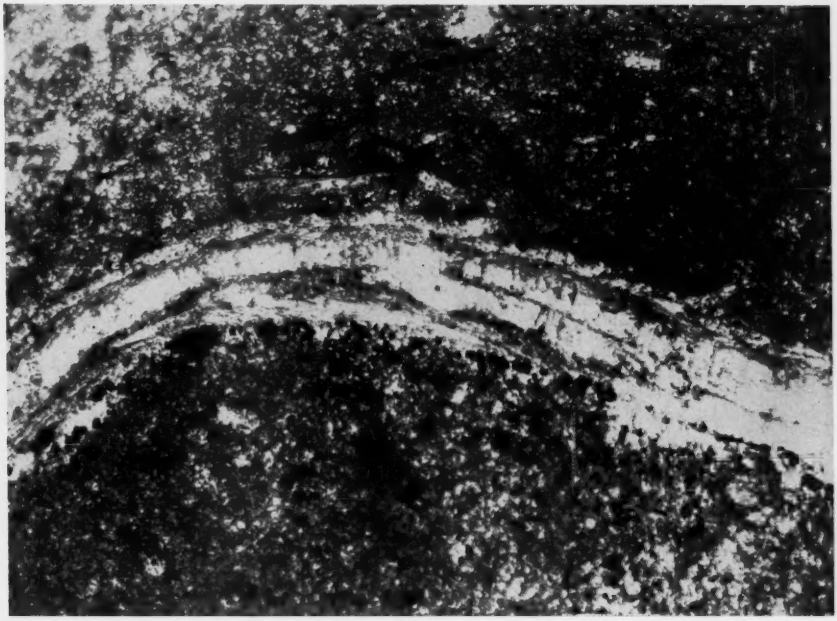

FIG. 6.—Worland Limestone. R. 29 W., T. 78 N., NE ¼, sec. 12. (Union tp., Dallas Co.) Concentration of pyrite along the ceiling of a brachiopod shell. Pyrite is also distributed throughout the rock. This limestone in slightly lignitic. Photomicrograph ×80.

organic matter is preserved and sulphides are formed below the interface (figs. 8 and 9).

DOLOMITE

Three modes of occurrence were noted:

1.—Granular rock-forming dolomite, constituting up to 90% of rocks heretofore considered limestones;
2.—Coarse crystalline dolomite, which has originated through recrystallization of shell material;
3.—Minute rhombs of ankeritic dolomite, distributed throughout the body of the rock.

It is the latter variety upon which we shall focus our attention. Upon weathering it yields limonite; commonly the rhombs have rusty edges, or they have decomposed to the point where all that remain are skeletal rhomb outlines stained with limonite (fig. 10). When fresh, this dolomite can generally be detected only by virtue of its distribution, size, and tendency for excellent rhombohedral form. In fact it was quite by accident that the writer realized after looking at many thin sections that these rhombs existed and with practice could be distinguished from the other constituents of the rock.

FIG. 4.—Myrick Station Limestone. R. 30 W., T. 79 N., SW ¼ sec. 25. (Jackson tp., Guthrie Co.) Broken, abraded fusulinids, slightly pyritized. The matrix is granular calcite; small foraminifer in center of photo is highly pyritized. Photomicrograph ×80.

FIG. 5.—Cooper Creek Limestone. R. 26 W., T. 75 N., SW ¼, sec. 19. (South tp., Madison Co.) Peculiar "mud ball" texture, with unusually high clastic silica (black) content. Hyaline calcite and granular calcite fill interstices. Pyrite (very white) is abundant, associated with the intersticial material, rarely with the matrix of the "mud balls." These structures are not organic. Fossil debris is scanty. Negative print ×8.

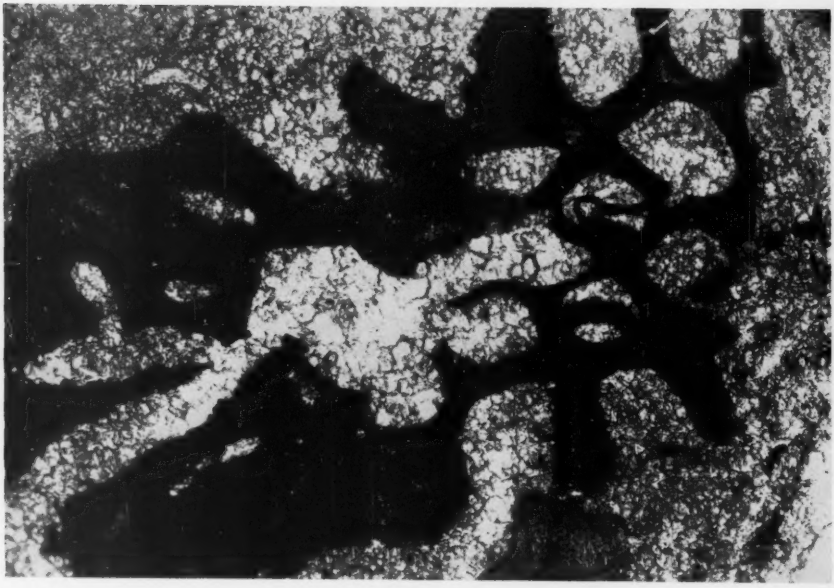

FIG. 7

FIG. 8

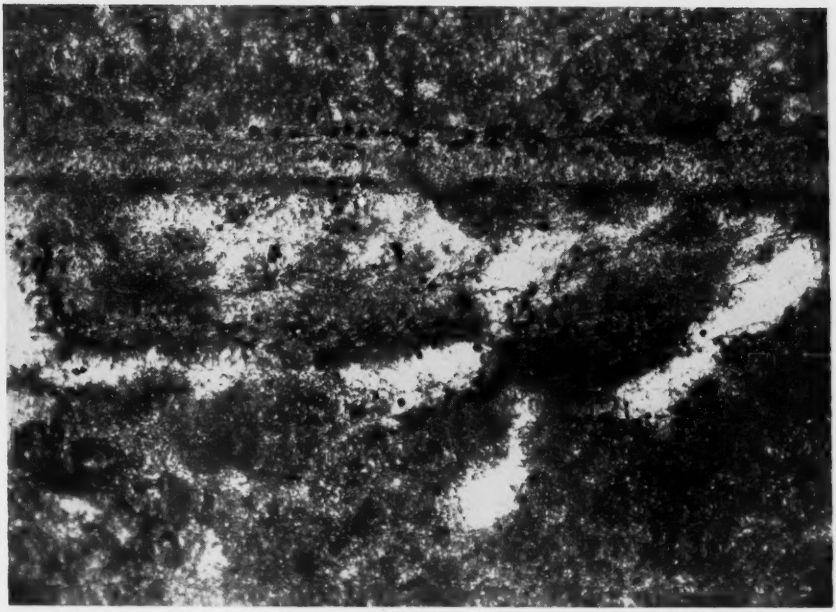

FIG. 9.—Enlarged view of a portion of a shell fragment of Fig. 8, revealing a concentration of pyrite along its margins; matrix is quite argillaceous. White areas represent partial recrystallization of the shell. Photomicrograph ×80.

It is significant that the magnesium of dolomite can normally be replaced only by iron in the ferrous state, since the ionic radius of Fe^{++} is not conformable to that of Mg^{++}. This implies that ankeritic dolomite should not have formed in the positive Eh environment above the depositional interface, but rather that it was a product of certain reactions in the reducing environment of the unconsolidated sediment below the depositional interface. Now it is known that organisms can cause small amounts of iron to be incorporated in the calcite lattice, and thus subsequent dolomitization of grains of such calcite would lead to the development of ankeritic dolomite. However, might not this take place more easily in the negative Eh environment where the iron tends to be in the reduced state? Whether formation of original ankeritic dolomite would occur in preference to the dolomitization of iron-bearing organic calcite is not known. Nevertheless, samples which contain ankeritic dolomite in their matrix exhibit no coarsening of texture as might be expected as a result of the dolomitization of a finely granular limestone. Moreover, the random distribution of individual minute dolomite rhombs—of the same size range as the granular calcite with which they are associated—is strongly indicative of crystallization of original ankeritic dolomite below the depositional interface more or less contemporaneous with pyrite.

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FIG. 7.—Myrick Station Limestone. R. 30 W., T. 79 N., SW ¼, sec. 25. (Jackson tp., Guthrie Co.) Highly pyritized fragment of a bryozoan; central portion may contain ankeritic dolomite. Fibrous structure of a brachiopod shell can be seen at bottom. Photomicrograph ×80.

FIG. 8.—Blackjack Creek Limestone. R. 26 W., T. 75 N. (Smith tp., Madison Co.) An irregular pod of argillaceous-fossiliferous, pyrite-bearing material surrounded by barren, relatively pure granular calcite. Pyrite (very white) is abundant in the "pod" as very tiny crystals, particularly in and around fossil fragments, some of which have been locally silicified. Negative print ×8.

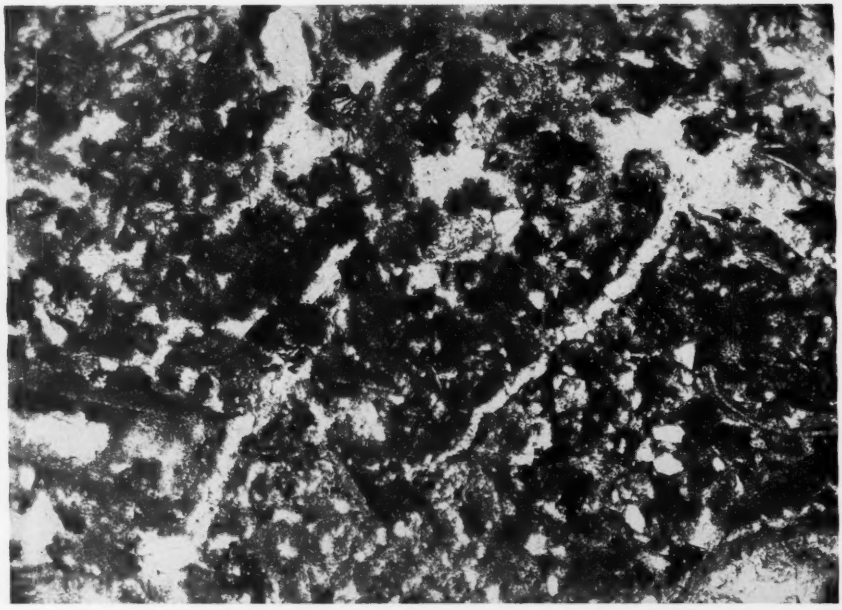

FIG. 10.—Ardmore Limestone. R. 29 W., T. 78 N., SW ¼, sec. 9. (Union tp., Dallas Co.) Ankeritic dolomite rhombs distributed throughout very fine-grained slightly argillaceous limestone; they are dark in color due to limonite staining. Photomicrograph ×80.

CLASTIC NATURE OF THE LIMESTONES

For years geologists generally accepted the idea that most limestones were wholly or in part the result of direct chemical precipitation. Recently, however, there has been a tendency to re-examine old evidence in the light of better knowledge of present-day processes and to conclude that mechanical activity has been in many cases the dominant factor in the final accumulation of calcareous rocks. Certainly the samples examined in this study appear to bear out this idea, and while they are not to be taken as representative of limestones in general, they have caused the writer to seriously doubt the importance of direct chemical activity in the accumulation of limestone.

EVIDENCE FOR MECHANICAL ACTIVITY

The most obvious manifestation of mechanical activity takes form in the variety of large abraded and broken fragments of shell material. Such debris is present in varying amount in almost all limestones (fig. 11). Not all fossil remains were broken, however. A few specimens were exceptional in that they were a sort of calcarenite (fig. 12) —subrounded grains of fossiliferous limestone in a scanty granular matrix—but these represent a special environment and do not have an important bearing on this discussion. In addition to the conspicuous coarse shell debris, there occurs in an overwhelming majority of cases very finely comminuted shell material. This often constitutes a significant portion of the matrix of the limestones, and is easily confused with the more usual granular calcite having no organic structure.

The granular matrix consists of individual crystals of calcite (and some dolomite) which are rounded to varying degrees. Thus they appear to have been precipitated chemically—or organically—and then to have been subjected to abrasion (fig. 1). Because of the solubility of calcite in an aqueous medium, it is most likely that this abrasion is nothing more than the result of partial solu-

tion of the grains. This process may be roughly divided into two categories: (1) corrosion of calcite crystals precipitated close to the surface of the water as they settle to the sea floor due to the decrease in pH and temperature with depth; and (2) mechanical winnowing of calcite grains on the sea floor by currents and very minor wave action. In this manner they are said to behave as clastics, although ultimately originating from chemical precipitation.

It will be objected that the variations in temperature and acidity with depth, in a shallow, relatively near-shore environment, are not sufficient to redissolve a portion of a calcite crystal, but it must be realized that very little solution is required to remove the edges from a rhomb of calcite 0.1 mm in diameter.

We may say then that these limestones have two main components: (1) true clastic fragments including comminuted shell material and, to a minor degree, fragments of pre-existing calcareous rocks; and (2) grains of calcite of chemical or bio-chemical origin, which from the time of their formation to their entombment in the sediment were subject to their physical environment just as any other type of clastic particle would be.

Calcite of definite chemical origin does exist in these limestones as coarse crystalline masses filling shells, spines, and other pore spaces, as well as that formed through the recrystallization of shell fragments. However, this type of calcite is certainly post-depositional—probably diagenetic, but possibly in part post-lithification.

CONCLUSIONS

In summary, it may be stated that: (1) Under favorable circumstances pyrite may form in bottom sediment beneath a marine open-circulation environment without constituting an impediment to benthonic fauna; (2) ankeritic dolomite may form in the same environment as, and contemporaneous with, pyrite; and (3) the main constituents of all the limestones studied, including the fine-

FIG. 11.—Exline Limestone. R. 26 W., T. 75 N., sec. 19. (South tp., Madison Co.) Heavy coating on brachiopod shell fragment. Pyrite as very tiny grains is disseminated throughout both the fragment and the matrix. Photomicrograph ×80.

FIG. 12.—Higginsville Limestone. R. 26 W., T. 75 N. (South tp., Madison Co.) Very fine-grained limestone, showing heavy Osagia crusts on brachiopod shell fragments. Hyaline calcite is common as pore fillings. Pyrite (very white) is distributed throughout. Negative print ×8.

grained calcite matrix, show evidence of mechanical activity.

ACKNOWLEDGMENTS

The writer is indebted to Dr. L. M. Cline for suggesting the problem which led to the writing of this paper and for invaluable assistance throughout the study. Thanks is also due Dr. S. A. Tyler for helpful discussion of many aspects of the problem.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 27, NO. 3, PP. 293–297
FIGS. 1–4, SEPTEMBER, 1957

COMPARISON OF PERCENTAGE AND RATIO DATA IN FACIES MAPPING[1]

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ABSTRACT

In facies map practice percentage maps and ratio maps are sometimes considered as strongly alternative ways of expressing certain facies patterns. The fixed relation between percentages and ratios for a common set of data implies that the maps are supplementary instead of competitive. It is shown in this paper that the conventional arc sine transformation of percentage data yields numbers identical with an arc tan transformation of the corresponding ratios, and that a single facies map of the transformed variables simultaneously presents some of the implications of both types of individual map. The transformation also facilitates further statistical analysis of the map data, especially in connection with analysis of variance.

INTRODUCTION

Percentage maps and ratio maps are common devices for showing areal variations in stratigraphic aspect. For a given area and stratigraphic unit composed of sand and shale, for example, the control points and measured thicknesses are the same for either a sand-shale ratio map or a sand percentage map. The decision to use one or the other depends on the specific kinds of information to be communicated by the map and on personal preference or assumed ease of interpretation.

The relation between the sand-shale ratio and percent of sand is such that when the ratio is $\frac{1}{4}$ the percentage is 20, when the ratio is 1 the percentage is 50, and when the ratio is 4 the percentage is 80. Thus, these three lines occupy the same positions on the two maps, subject to some differences due to the method of contour interpolation. It is the purpose of this paper to explore some of the relations between ratio and percentage maps in terms of differences in the way they communicate their information, and to show that it is possible to apply certain transformations to the data such that both the ratios and percentages yield identical values.

RELATION BETWEEN RATIOS AND PERCENTAGES

For a two-component system, such as sand and shale, where A represents the thickness of sand and B represents the thickness of shale at a control point, the sand-shale ratio is $R=A/B$ and the sand percentage is $P=100A/(A+B)$. By algebraic substitution it may be shown that:

$$\left.\begin{aligned} R&=P/(100-P)\\ P&=100R/(1+R)\end{aligned}\right\} \qquad (1)$$

The relations in equation (1) are shown graphically in figure 1. The upper portion shows arithmetic spacing of the percentage values, with the corresponding values of geometrically spaced ratios set opposite them. The lower portion shows the geometric ratios laid out on an implied logarithmic scale, with the corresponding percentages below the line. When percentages are contoured arithmetically, the effect is as though the corresponding ratio lines near $R=1$ were "stretched out." When the ratios are contoured on geometric intervals, the percentage lines near $P=50$ are "compressed."

Figure 1 applies to any ratio-percentage combination based on a common set of data, even when more than two end members are involved. The clastic ratio, for example, may be set up by letting the sum of the thicknesses of sand and shale equal A, and letting the sum of the thicknesses of the non-clastics equal B. Then the clastic ratio is A/B and the percentage of clastics is $100A/(A+B)$, and the relations of equation (1) apply.

It is largely the stretching and compressing effects that produce differences in ratio and percentage maps. These effects may be desirable when certain kinds of gradients are to be emphasized on the maps. In the more general case, however, where no particular subjective emphasis is intended, it may be preferable to use such a trans-

[1] Manuscript received April 16, 1957.

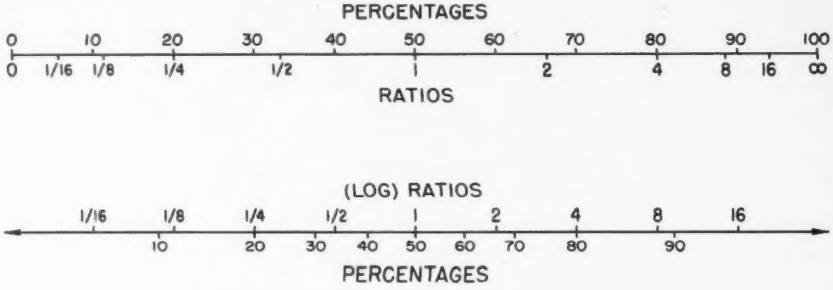

FIG. 1.—Relations between percentages and ratios, showing effects of arithmetic percentage spacing (top) and logarithmic ratio spacing (bottom).

formation of the percentage or ratio data that identically the same map is obtained from either. This may be accomplished for percentage data by a well-known conventional procedure.

THE ARC SINE TRANSFORMATION FOR PERCENTAGE DATA

Although one intuitively contours percentage data on an arithmetic interval, there is some implication that a change from 50 to 55 percent, say, is not as significant as a change from 5 to 10 percent. That is, one may question whether equal differences along a percentage scale are necessarily of equal significance in interpretation. If not, some contour system other than the conventional equal interval spacing may be appropriate.

When items are counted (instead of being measured as on an electric log), the number X_1 of items having a given attribute in a count of N items may be expressed as the proportion X_1/N, or as the percentage $100X_1/N$. In either case the proportions or percentages are binomially distributed (Bennett and Franklin, 1954, p. 111–114), providing that X_1 is not very small. Binomially distributed data have a variance dependent on N, so that for some kinds of statistical analysis a transformation is desirable to stabilize the variance. The arc sine square root transformation to an angle in degrees does this, yielding a constant variance of $821/N$ (Kempthorne, 1952, p. 155). Tables for converting percentages to their arc sines in degrees are given in Snedecor (1946, p. 449).

Stratigraphic data are not based directly on counts of items (except in heavy mineral maps and similar devices based on counts), but conventional percentage data do tend to be distributed binomially. The arc sine transformation is applicable also to such data, although the variance probably contains some non-binomial components. The effect of the arc sine square root transformation is to make the change from 50 to 55 percent much less important than a change from 5 to 10 percent, but more important than a change from 5 to 6 percent (Krumbein and Tukey, 1956, p. 335). That is, by transforming percentages to their arc sine equivalents, and contouring on an equal arc sine interval, a "weighting factor" that appears to be appropriate for percentage data is provided. As will be shown in the following section, there is an equivalent transformation for ratio data that gives exactly the same value for the transformed ratio as the arc sine does for the corresponding percent.

The arc sine square root transformation for percentage data is obtained by expressing the percentage as a proportion, extracting its square root, and finding the angle θ whose sine is equal to the square root. Thus, for the percentage 33.3% the proportion is 0.333. The square root of this is 0.577. The angle θ whose sine is 0.577 is 35.2°. The number 35.2 is the arc sine square root transform of 33.3%.

THE ARC TAN TRANSFORMATION FOR RATIO DATA

When the ratio corresponding to a given percentage value has its square root extracted, and if this square root is looked up

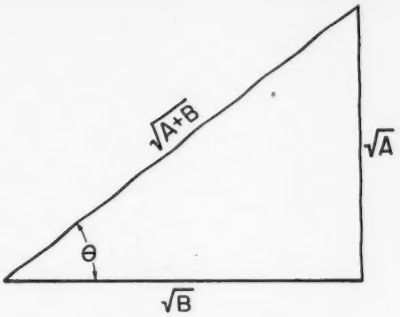

FIG. 2.—Right-angle triangle showing lengths of sides.

in a table of tangents, the angle θ whose tangent is equal to the square root will be exactly the same as the angle θ obtained when the square root of the original percentage value (expressed as a proportion) is transformed to its arc sine. As an example, the ratio corresponding to 33.3% is 0.500, by equation (1). The square root of 0.500 is 0.707, and the angle θ whose tangent is 0.707 is 35.2°. Thus the value 35.2 is simultaneously the arc sine value of the proportion root and the arc tan value of the corresponding ratio root.

The identity can be demonstrated trigonometrically. Let $R=A/B$ and $p=A/(A+B)$. Here the percentage is expressed as its proportion, p. The problem is to demonstrate that arc tan $\sqrt{R}$ = arc sin $\sqrt{p}$. Figure 2 shows a right angle triangle with sides $\sqrt{A}$ and $\sqrt{B}$, from which it follows that the hypothenuse is $\sqrt{A+B}$. In terms of the angle θ, $\sin\theta=\sqrt{A/(A+B)}$ and $\tan\theta=\sqrt{A/B}$. Hence, θ = arc sin $\sqrt{A/(A+B)}$ = arc tan $\sqrt{A/B}$, and the relation is proved. In tables of the arc sine transformation p is usually expressed as the percentage, $P=100\,p$.

The writer is not aware of any published tables of arc tan $\sqrt{R}$, but because the arc tan transformed value for any R is exactly the same as the arc sine transformation for the corresponding p, available tables of the arc sine transformation may be used for ratios by first converting them to their percentages with equation (1). When only a few ratios are involved, the transformation may be made directly with tables of tangents.

The effect of the transformation is shown in figure 3, where the arc tan $\sqrt{R}$ = arc sin $\sqrt{p}$ values are shown on an arithmetic scale, with the percentages value indicated above and the ratio values indicated below. As may be seen, the effect is to compress the percentage values toward the center as in the lower diagram of figure 1, and to stretch the ratio values near the center as in the upper diagram of figure 1, but the degree of compression or stretching may be reduced. It is also apparent that the percentage and ratio scales near the center of figure 2 are much the same from about 30 to 70 percent and from $\frac{1}{2}$ to 2 (more nearly $\frac{1}{3}$ to 3). It is to be recalled that the percentage values in the diagram are computed as proportions, and then multiplied by 100 to express them as percentages.

As an illustration of these relations, the maps of figure 4 are presented to show how the same data plot in the three map forms. The data are artificial, but they are based on situations not uncommon in facies mapping, where a wide variation in sand content occurs in a stratigraphic unit. The maps show that the percentage lines are stretched out at the extremes (between 90 and 100%), and that the ratio lines are compressed at the extremes (between ∞ and 8) as brought out in figures 1 and 3. The arc sine = arc tan map is contoured on equal θ values. This map may be called an "angular sand map" or a "theta sand map."

The inset graph with the maps shows

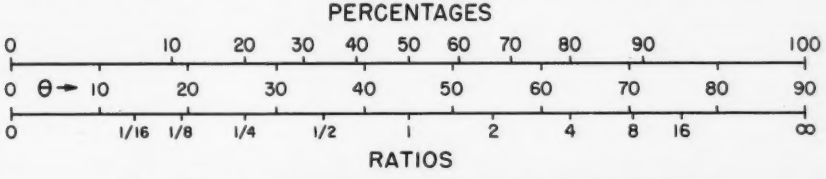

FIG. 3.—Relations between percentages and ratios in terms of arithmetic spacing of transformed values (θ).

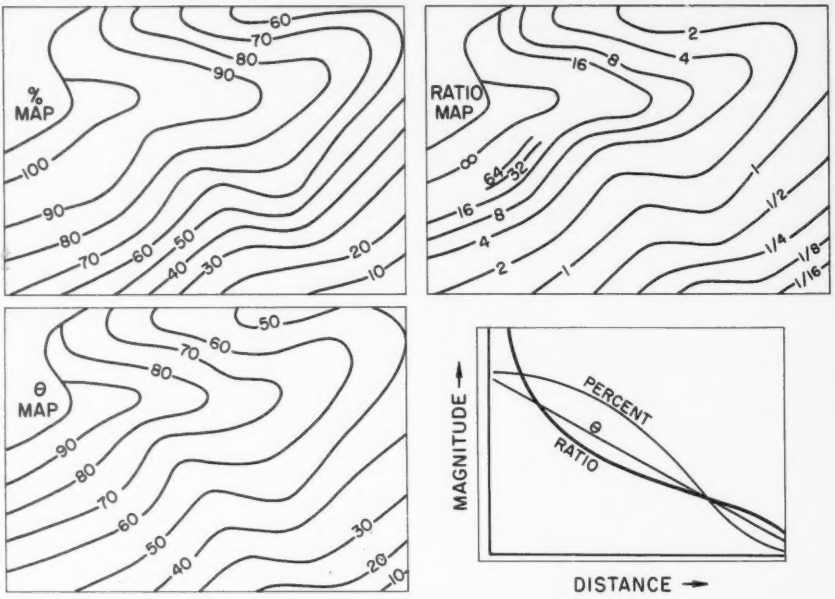

FIG. 4.—Comparison among percentage, ratio, and "theta" map for hypothetical stratigraphic section composed of sand and shale. The lower right graph schematically shows the slopes across the southwestern part of the maps.

diagrammatically how the profiles compare across the southwestern part of the map, where the θ lines have a nearly uniform slope. The percentage profile is bowed convexly, and the ratio profile is essentially a negative exponential along its left half.

STATISTICAL IMPLICATIONS OF THE ARC SINE AND ARC TAN TRANSFORMATIONS

For count data the arc sine transformation to θ stabilizes the variance of percentage data as mentioned earlier. In the same manner the arc tan stabilizes the variance of the corresponding ratio data. The importance of this is that some kinds of statistical analysis of the facies data are facilitated by the transformation. In analysis of variance a basic assumption is that the variance is homogeneous (Dixon and Massey, 1957, p. 179). The transformation assures this result in count data, and appreciably improves the situation in percentage data based on measurement of stratigraphic sections.

The arc tan transformation to θ also serves an additional useful purpose with ratio data. The occurrence of infinity values, for example, hinders the computation of means inasmuch as any set of numbers with an infinity value adds up to infinity. The angle θ ranges from 0° to 90°, so that the values remain well within the finite range. The setting of confidence limits about mean values is also facilitated inasmuch as the limits are symmetrical about the arc sine and arc tan mean, but they may not be so when computation is made with the original variates.

Although the arc sine transformation generally improves statistical analysis of percentage data, percentages in the range from about 30 to 70% can be treated without transformation, and the same applies to ratio data over the range from about 0.3 to 3.0. Furthermore, ratio data over the range from perhaps 8 to $\frac{1}{8}$ can be approximately handled with a logarithmic transformation. When the range of the observational data is large, as in the maps of figure

4, the best over-all transformation for statistical analysis appears to be the arc sine or arc tan.

Another potential advantage of the arc sine and arc tan transformations is brought out by figure 2. The orthogonal relation between $\sqrt{A}$ and $\sqrt{B}$ permits application of complex variables to facies mapping. For example, the co-ordinates of the point at the upper right corner of the triangle can be expressed as $(\sqrt{B}+i\sqrt{A})$, where $i=\sqrt{-1}$. Each control point may then be considered as a vector of length $r=\sqrt{A+B}$ and direction θ. The vector thus combines thickness and facies data, and permits development of single contour system maps that simultaneously convey thickness and compositional data. This interesting extension of facies mapping is being examined in greater detail for later presentation and illustration.

CONCLUDING REMARKS

Although it is emphasized here that the arc sine and arc tan transformations facilitate statistical analysis of percentage and ratio data, and provide a single contour system map for ratios or proportions, the writer in no way wishes to imply that percentage and ratio maps be discarded. These maps, in fact, permit emphasis of certain kinds of rates of change which the map maker may wish to bring out, as for example emphasis on exponential trends in ratio maps. The choice of method between percentage, ratio, or θ still depends on the kinds of information to be communicated by the map, as well as on the usefulness of the map in predicting facies relations among and beyond the control points.

Extensions of facies mapping techniques have the principal value that they enlarge the choice of methods for presenting stratigraphic data. The large subjective element in map preparation and interpretation suggests that the worker have as wide a choice of methods as possible, to permit the kind of presentation that careful study of the stratigraphic unit suggests to him. In the writer's opinion most mapping methods are supplementary rather than competitive, and each conveys its own particular shades of meaning.

The writer acknowledges his indebtedness to his colleagues at Northwestern University, especially Dr. H. A. Slack, and to numerous geologists in the oil industry for helpful suggestions and discussion of the principles presented in this paper. Support of this work by funds from the Graduate School, Northwestern University, is gratefully acknowledged.

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FIG. 1, PLATE 1, SEPTEMBER, 1957

AUTHIGENIC MONTMORILLONOID CEMENT IN ANDESITIC SANDSTONES OF CENTRAL CALIFORNIA[1]

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ABSTRACT

Distinctively blue upper Tertiary sandstones, composed almost entirely of andesitic detritus derived from upper Tertiary andesite breccias of the northern Sierra Nevada, are widespread in central California. These volcanic arenites are well sorted and loosely cemented by a thin coating of authigenic clay mineral encasing each detrital grain. The clay has grown in laths or fibers oriented perpendicular to the surfaces of the clastic grains. The blue color is due to the effect of this translucent coating on light reflected from the dominant dark-colored grains beneath it. Optical, X-ray, differential thermal, and chemical data indicate that the mineral is a montmorillonoid intermediate between beidellite and nontronite, a composition not commonly recorded in published analyses. The clay mineral crystallized from pore solutions permeating the sandstone, the material probably being mostly derived by solution of the groundmass of the andesite rock fragments, which make up half or more of the sandstone.

INTRODUCTION

Sandstones having a distinctly blue color are widespread in central California, but their occurrence is limited to Tertiary strata composed largely of andesitic debris. The blue aspect of these rocks, which has long attracted the attention of geologists working in the area, is caused by a thin mineral coating that encases the sand grains and pebbles. In the first recorded attempt to determine its composition, Arnold and Anderson (1910, p. 97) concluded that this coating is vivianite, but Anderson and Pack later recanted and decided that it is opal or chalcedony (1915, p. 82). Since then, this material has commonly been referred to as opaline (Huey, 1948, p. 43; Wilson, 1943, p. 241). Bramlette (1934, p. 1574) judged the coating to be chloritic, and the present study shows that it is a montmorillonoid having a composition intermediate between beidellite and nontronite.

GENERAL OCCURRENCE OF ANDESITIC SANDSTONES

The principal source of Californian andesitic sandstones, in which the montmorillonoid coatings have formed, was the Sierra Nevada north of Yosemite National Park, where a great volume of andesite was extruded during the upper Miocene and Pliocene epochs. As a result of these eruptions most of the northern Sierra Nevada was blanketed by a thick accumulation of andesite. The locus of the eruptions was near the crest of the range, but a great quantity of andesitic debris was swept down the westward slopes and deposited on broad alluvial plains and in shallow bordering seas which lay in the area where the Central Valley and Coast Ranges now stand.

Today, andesitic sandstone blankets the western foothills of the central and northern

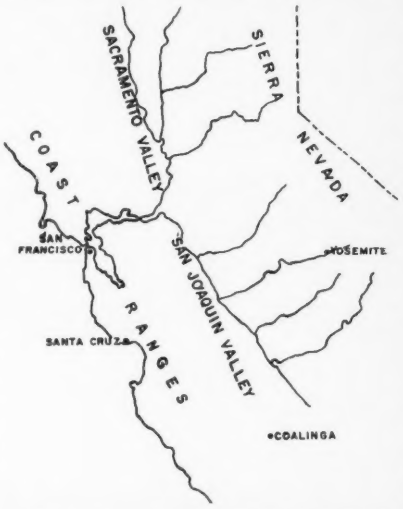

FIG. 1.—Map of central California indicating the area under consideration and locations of places mentioned in the text.

[1] Manuscript received August 29, 1956.

TABLE 1.—*Correlation chart of blue sandstone formations in central California*

AGE	Central Sierra Nevada	San Francisco Bay Area	Santa Cruz Area	Coalinga area
PLIO-PLEISTOCENE	?			San Joaquin Formation
PLIOCENE	?		Purisima Formation	Etchegoin Formation
MIO-PLIOCENE	Mehrten Formation	Neroly Formation		
		Cierbo Formation		

Sierra Nevada (Mehrten Formation) and extends westward and southwestward beneath the alluvium of the Sacramento-San Joaquin Valleys, as shown by numerous well cores. Farther west in the folded central Coast Ranges it crops out in many places over a distance of about 200 miles from the San Francisco Bay region on the north to the Coalinga area on the south. The principal formations containing andesitic sandstone and their general locations are summarized in table 1 and in figure 1. In some parts of the Coast Ranges, deposits of andesitic sandstone are as much as 2000 feet thick.

The andesitic debris transported westward from the northern Sierra Nevada was at times mixed with detritus derived from non-volcanic sources. During the upper Miocene and Pliocene epochs, however, the flood of volcanic detritus was so great that widespread deposits of nearly pure andesitic debris were laid down. Table 2 summarizes the mineral composition of representative samples from various of the formations; in general, one can take the combined percentage of quartz and non-volcanic rock fragments as an approximation of the contamination from non-volcanic sources.

PROPERTIES OF THE AUTHIGENIC MONTMORILLONOID CEMENT

General Occurrence and Habit

Authigenic clay, which is shown by X-ray, optical, and differential thermal data to be a montmorillonoid, is normally the only cement in the blue sandstones (pl. 1 A). It occurs on grain surfaces bordering original pore spaces and has not been noticeably crushed by any subsequent compaction. The sandstones are well-sorted, friable deposits composed of subangular grains. Andesite rock fragments make up about one-half of the detrital material (table 2), and most of the remainder is intermediate plagioclase and ferro-magnesian minerals commonly found in andesites, the latter usually comprising 10 to 25% of the rock. Quartz seldom exceeds 15% and is often nearly absent. In the terminology of Williams, Turner, and Gilbert (1954, p. 290–294; fig. 103, p. 309) these rocks are andesite arenites, a variety of lithic arenite (cf. pl. 1).

The montmorillonoid coating has a maximum thickness of 0.02 mm. and is made up of fibers only .001 to .003 mm. across. Despite its thinness, the coating is distinctly birefringent and is always composed of laths or fibers of clay mineral oriented with their long dimension normal to the surface of the detrital grain on which the coating developed. Some of the thicker coatings consist of two layers, or generations, of fibers having slightly different refractive indices, birefringence, and color. These features, illustrated in plate 1, C and D, suggest that the montmorillonoid was precipitated slowly on the surfaces of detrital grains from solutions in the pores of well-sorted sandstones.

The blue aspect of many of the coated sandstones is a superficial optical effect which appears only where the coating is very thin, less than about 0.01 mm. Such thin coatings are translucent, and the blue

TABLE 2.—*Compositions of typical andesitic sandstones containing authigenic montmorillonoid cement*

Formation	Mehrten	Cierbo	Neroly	Etchegoin	San Joaquin	Purisima
Sample No.	169S-20	169S-11	169S-76f	169S-58a	169S-52a	169S-108
	%	%	%	%	%	%
Andesite fragments	54	48	53	45	51	43
Intermediate plagioclase	20	13	26	17	11	7
Potash feldspar	3	6	4	3	2	6
Quartz	<1	11	3	7	12	14
Other rock fragments	<1	7	<1	7	6	13
Total heavy minerals	23	15	14	21	18	17
Hornblende	4	7	<1	8	6	5
Hypersthene	6	5	5	4	4	5
Augite	6	3	5	5	4	4
Iron oxides	7	<1	4	4	4	3

color results because light reflected from underlying dark-colored grains has passed through the translucent coating. That the color is produced in this general way is indicated by the effect of moisture, for the blue color appears distinctly only when the rock is dry. When it is wet, the coating seems transparent and the rock takes on the color of the detrital fragments, which are mostly dark gray or black. Light-colored grains such as feldspar appear milky white or buff when coated by montmorillonoid, but because the predominant fragments in these sandstones are dark colored, the rock as a whole takes on a bluish cast. The thin coatings adhere very closely to the grains. They may be scraped from pebbles with the fingernail, and behave much like a thin layer of wax.

Not all coated sandstones appear blue. When the coating is more than about 0.01 mm thick, it becomes opaque and obscures the underlying grains. The rock then has the color of the coating itself and is usually tan unless stained by iron oxides. Thus the thick-coated brown sandstones are not so conspicuous as the thinner-coated blue sandstones. In addition, the thick coatings are easily disrupted from the grains, and even careful fracturing of the rock may so loosen them that upon casual examination the material appears to be interstitial.

Optical Character

The color of light transmitted through the coating is usually pale tan to greenish tan. The montmorillonoid fibers are either length-fast or length-slow but always have the same orientation throughout any one coating, although a few dual-layer coatings have one layer of fibers which are length-fast and the other layer of fibers which are length-slow. The birefringence is low, and the material so finely crystalline that parts of some coatings appear isotropic.

Montmorillonoids, other than some iron-rich nontronites, are optically negative and have small optic axial angles, so that the

→

EXPLANATION OF PLATE 1

A.—Neroly blue andesitic sandstone. Note detrital grains encased by montmorillonoid coatings which have been partly disrupted to expose dark minerals or rock fragments beneath. ×10.

B.—Mehrten blue andesitic sandstone weakly cemented by thin authigenic montmorillonoid coatings around the detrital grains. Note hypersthene near center with serrated terminations showing solution from beneath the coating which still preserves the outline of the original grain; augite (far right and lower right), and hornblende (upper right and lower left) are unetched. ×66.

C.—Detail of authigenic montmorillonoid cement coating feldspar (upper right), andesite rock fragment (lower right), and rimming large pore space in left center. Note two generations of coating with outer rim of iron oxide. ×320.

D.—Same as C with crossed nicols. Note difference in birefringence of two generations of coating.

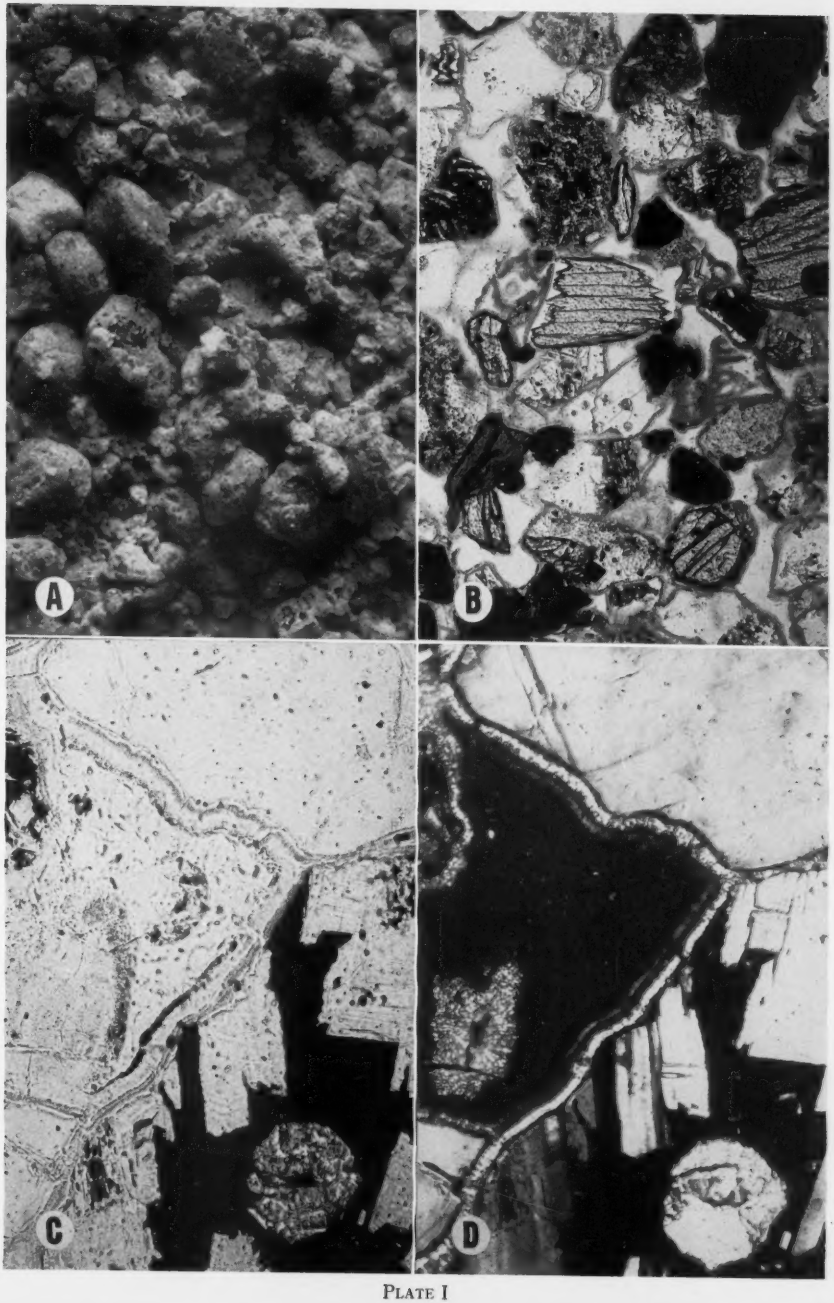

PLATE I

value of the beta index is close to that of gamma (Winchell, 1945, p. 516). If this is correct for the montmorillonoid here being described, fibers that are length-slow have a refractive index parallel to the length which is near gamma. If these fibers are randomly oriented about the long axis, the index perpendicular to the lengths of a group of parallel fibers is then approximately $(\gamma+\alpha)/2$ since gamma and beta are nearly equal. The greatest birefringence observed for such a group would approximate $\gamma-[(\gamma+\alpha)/2]$ or $(\gamma-\alpha)/2$. If alpha is approximately parallel to the long axis of the length-fast fibers, the highest birefringence observed in coatings with this optical orientation would be $[(\gamma+\beta)/2]-\alpha$, or nearly $\gamma-\alpha$. The larger the optic angle the greater the error in these approximations. Birefringences of four samples each of length-slow and length-fast fibers are shown in table 3. These were calculated using a Berek compensator and thicknesses measured with a calibrated micrometer focusing adjustment on the microscope. The birefringences found are lower than those usually recorded for montmorillonite and beidellite, but are of the same order as those found by some workers (Winchell, 1945, p. 516). The writer is puzzled by the large range in birefringence recorded in the literature for montmorillonoids of approximately the same composition.

The air dry material at room temperature has both maximum and minimum indices within the range from 1.50 to 1.54, and the basal spacing of the material is about 15 kx, indicating two layers of inter-lattice water. The range in indices observed in a single sample makes it impossible to calculate the birefringence accurately from these readings. Refractive indices were measured with immersion oils that are completely insoluble in water and have low volatility.[2] Oils without these two properties do not yield consistent results on expanding-lattice clay minerals, for slight water solubility of the immersion oil will cause some displacement of the interlayer water, and a resultant change in the refractive index (Vendel, 1945).

[2] The compounds used as end members for mixed oils were n-tetradecane (1.426), mineral oil (1.484), and alpha-chloronaphthalene (1.633). The last is not quite as water insoluble as could be desired, but is the best of the common higher index compounds.

X-ray Data

X-ray patterns show the coating to be an expanding layer-lattice mineral belonging to the montmorillonite group (Ross and Hendricks, 1945). Following the suggestion of MacEwan (1951, p. 86), such a mineral is here called a montmorillonoid. The term montmorillonite is reserved for a member of more limited composition in which the Si:Al ratio is about $2\frac{1}{2}$:1 and some Mg and usually some Fe occurs in octahedral coordination. The name beidellite is used for a somewhat more aluminous member having a Si:Al ratio of $1\frac{1}{2}$:1 or less. Nontronite is a ferruginous montmorillonoid having Fe as the most abundant cation in octahedral position. There is probably a complete range of substitution of Fe for Al from beidellite to nontronite, subject to certain limitations (MacEwan, 1951, p. 98), but analyses published by Ross and Hendricks (1945, p. 34, 35) show few with between one and two Fe atoms per unit cell out of a total of approximately four filled octahedral positions. As a result, the beidellite-nontronite boundary was provisionally taken by MacEwan (1951, p. 99) to be "somewhere in this gap."

TABLE 3.—*Optical properties of authigenic montmorillonoid cement coating*

Sample No.	Orientation	$\gamma'-[(\beta+\alpha)/2]$ $\sim(\gamma-\alpha)/2$	$\gamma-\alpha$ (approx.)
2a	Length-slow	0.005	0.010
2c	Length-slow	0.003	0.006
11	Length-slow	0.0005	0.001
14	Length-slow	0.005	0.010
128*	Length-slow	0.0035	0.007

Sample No.	Orientation	$[(\beta+\gamma)/2]-\alpha$ $\sim\gamma-\alpha$
52a	Length-fast	0.011
52b	Length-fast	0.015
65a	Length-fast	0.011
97	Length-fast	0.016
128*	Length-fast	0.008

* Two-layered coating with different orientation in each layer.

Substitution of Fe and Mg for Al in octahedral position causes an increase in lattice dimension. Since the 060 reflection is usually fairly sharp and strong in montmorillonoid patterns this spacing may be used to estimate the degree of such substitution. MacEwan (1951, p. 106, fig. IV, 8) produces a graph showing the change in the b_0 dimension with substitution of Fe for Al in the beidellite-nontronite series, between 4.05 and 4.22 filled octahedral positions per unit cell. (These are the limits found by Ross and Hendricks.) The 060 spacing of the montmorillonoid material here being considered is about 1.505, corresponding to a b_0 of 9.03, and according to MacEwan's graph this would mean somewhere between 1.4 and 1.8 Fe atoms per unit cell depending upon the total population of the octahedral layer. This suggests that the composition of this authigenic montmorillonoid lies in the supposed gap between beidellite and nontronite.

It is interesting to note that of 40 montmorillonites and beidellites listed by Ross and Hendricks, 27 are bentonites presumably derived from the alteration of tuffs; 6 are pegmatite clays; 4 are gouge clays; 2 are from soils; and 1 is a vesicle filling. Of 8 nontronites for which they give this information, 7 occur as alteration zones or veinlets in metamorphic rocks, 1 is material from between two basalt flows. Only the montmorillonites from soils and in vesicles, and the nontronite from basalt seem likely to have originated by free precipitation from solution in a fashion at all similar to the montmorillonoids here being investigated.

Differential Thermal Data

Differential thermal curves of two samples of the coating have strong low-temperature endothermal peaks representing loss of interlayer water, broad low endothermal saddles extending from about 500 to 700 degrees caused by the loss of internal OH-water, and endothermal-exothermal couples between 800 and 900 degrees. The loss of internal water at intermediate temperatures in the montmorillonoids provides some evidence as to composition. The octahedral Fe-OH bond is evidently weaker than the Al-OH bond with the result that the endothermal peak representing the breaking of this bond and the loss of OH-water occurs at about 500 degrees in nontronite, and 700 degrees in montmorillonite and beidellite. Thus, the broad saddle extending between 500 degrees and 700 degrees given by the montmorillonoid coatings may be the result of a considerable amount of both Fe and Al in octahedral co-ordination, which causes loss of water to extend over a considerable temperature range.

Chemical Data

Attempts to determine the precise quantitative chemical composition of the montmorillonoid coatings were unsuccessful. Samples of reliable purity could not be prepared, but several that were examined by flame photometric and spectrographic methods suggest that the Fe_2O_3 content is in the neighborhood of 10–15% and the MgO content less than 1%, in agreement with the other analytical data.

CONCLUSIONS

The following general conclusions seem to be indicated by available evidence:

1.—The authigenic cement under consideration is montmorillonoid and probably has a composition intermediate between beidellite and nontronite. Such a composition is rarely recorded in published analyses, but it may perhaps be explained by the probable crystallization of this montmorillonoid from free pore solutions at low temperature.

2.—The montmorillonoid cement has been precipitated on and has grown outward from the surfaces of detrital grains along open pores in permeable sandstone.

3.—Montmorillonoid coatings are found only in andesitic sandstones, and they are equally abundant both far from and close to the volcanic source. This suggests that the substance of the montmorillonoid was derived by solution of some material within the sandstone itself. The one detrital component common to all the montmorillonoid-coated sandstones studied is fragments of intermediate volcanic rock, and probably the cryptocrystalline and glassy groundmass of these has been the chief source of the montmorillonoid material. The evidence for this is not conclusive, however; many of these rock fragments, and also scattered

glass shards, remain unaltered in coated sandstones, but others have clearly been converted to a montmorillonoid clay.

Bramlette (1934, p. 1574) once suggested that the coating in andesitic sandstones was related to the occurrence of detrital hypersthene. This mineral, together with hornblende and augite, is abundant in most of the coated sandstones, but it does not occur in all of them. Commonly both hypersthene and augite have been partly dissolved, as shown by their serrated (etched) terminations, but this solution often has occurred beneath the montmorillonoid coating which preserves the outlines of original grains (pl. 1 B).

4.—Authigenic montmorillonoid has not been observed either in sandstones completely cemented by calcite, or in concretions found in montmorillonoid-coated sandstones. This fact suggests that such calcite was precipitated early and sealed the rock before the formation of montmorillonoid cement. Some coated sandstones, however, do contain scattered calcite which formed later than the montmorillonoid; in these the calcite fills some of the interstices on top of the montmorillonoid coating.

5.—Montmorillonoid cement is equally common in both marine and non-marine sandstones which suggests that environment of deposition and connate waters did not control formation of this cement. It probably was not a very early diagenetic product. Nor is it related to the present surface and vadose waters of today, because it occurs in deep well cores beneath the Central Valley where the sandstone has not been uncovered since deposition. In the severely folded andesitic sandstones of the Coast Ranges there seems to be no relationship between structure and the occurrence of the montmorillonoid, which perhaps means that it formed before the intense Plio-Pleistocene Coast Range deformation.

In this connection, Doell (1956) has recently published some studies of remanent magnetization in the blue andesitic sandstones of the Neroly Formation in the Central Coast Ranges. He attributes the magnetization in these rocks to iron oxides crystallized in the interstices of the sandstone, and he concludes that they crystallized after folding. The present author has observed such iron oxides as films on top of montmorillonoid coatings (pl. 1 C).

The requisites for the development of the montmorillonoid, therefore, seem to be (1) andesitic material, especially rock fragments, (2) high permeability, (3) intrastratal solutions saturating the pores and probably migrating slowly, and (4) a considerable length of time. It seems very unlikely that these conditions have not been commonly met elsewhere, especially in the circum-Pacific belt of Tertiary andesitic volcanism. Montmorillonoid cemented sandstones should be more common than one would judge them to be from the literature.

Andesitic sandstones having montmorillonoid coatings occur outside the area discussed in this paper. Several samples collected by Mr. John T. Alfors in the Pancake Range in eastern Nevada were examined by the author and found to be essentially identical to the Californian sandstones. Ross, Miser, and Stephenson (1929, p. 184) also report a somewhat similar occurrence, believed by them to be glauconite or celadonite, in late Cretaceous phonolitic sandstones near the Texas-Oklahoma-Arkansas boundary.

ACKNOWLEDGMENTS

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 27, NO. 3, PP. 306–312
FIGS. 1–5, SEPTEMBER, 1957

THREE-DIMENSIONAL SHAPE ANALYSIS OF FINE-GRAINED SEDIMENTS[1]

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Birmingham University, England

ABSTRACT

A new method of three-dimensional shape measurement of silt particles is presented and methods of analysing the parameters obtained are discussed. The sample is dispersed, placed on a microscope slide, and shadowed with gold *in vacuo*. Spherical granules are also placed on the slide to enable the shadow angle to be obtained. Several areas on the slide are photographed at large magnifications and measurements made from these photographs under a binocular microscope, the actual measurement being performed at total magnifications of between 3000× and 10,000×. Three parameters are found for each particle. Two are measured directly from the photograph and the other calculated from the length of the shadow. The parameters obtained ($a>b>c$) are plotted as a point for each particle on the chart b/a against c/b. Histograms of shape and sphericity may be derived from this chart.

INTRODUCTION

The shape of a particle is a fundamental characteristic which is at least as important as size and density to the sedimentary petrologist, but, because of the difficulty of measuring the shape and especially of expressing shape in a form which enables sediments to be compared, it is a character which has been largely ignored in the description of argillaceous rock types. Two-dimensional methods of shape analysis have long been the only approach; and the best of these is by Pye and Pye (1943) which, for a sand, does give results consistent with true sphericity. Clays and silts, which are the type of sediment studied by the author, are not amenable to this type of analysis as it assumes that the grains are more or less equidimensional. It may be true for sands, but even so, Pye and Pye end their paper by commenting that "there is a great need of a method of securing a truer measure of the sphericity of grains. This method will probably have to depend upon a three-dimensional analysis." This has, in fact, now been done. Huitt (1954) published a short note which described the illumination of sand grains by ordinary light, but he does not mention his method of calculating the height from the shadow-length. This has been superseded by the work of Aschenbrenner (1955) on sand grains. He uses a photogrammetric method involving a specially made stereoscopic camera, with which measurements are made at a total magnification of 40×. It is felt that the difficulty of taking stereo-photomicrographs of very small particles makes this method impracticable for use with silts and clays. (But see Calbick, 1950.)

THE METHOD

The shadowing technique employed in electron microscopy was used to find the particle height. The method of analysis developed is as follows.

The sample to be analysed is dispersed in water, using a peptiser if necessary. It is diluted until the suspension appears colourless, a drop is then put onto a glass microscope slide by means of a pipette, and the water allowed to evaporate. The concentration of particles is checked at all stages under the microscope. The particles have to be sufficiently numerous to include a hundred or more in one field of view (at 150×) without being so close that their shadows mutually interfere.

It is also necessary to place some spherical particles onto the slide to enable the shadow angle to be calculated. Kallodoc Acrylic Granules, supplied by Imperial Chemical Industries Ltd. (Plastics Division), are suitable; they are about 20–30 microns in diameter.

The slide is then put into the shadowing chamber so that the glancing angle of the rays from the source is about 20°. The best

[1] Manuscript received August 31, 1956.

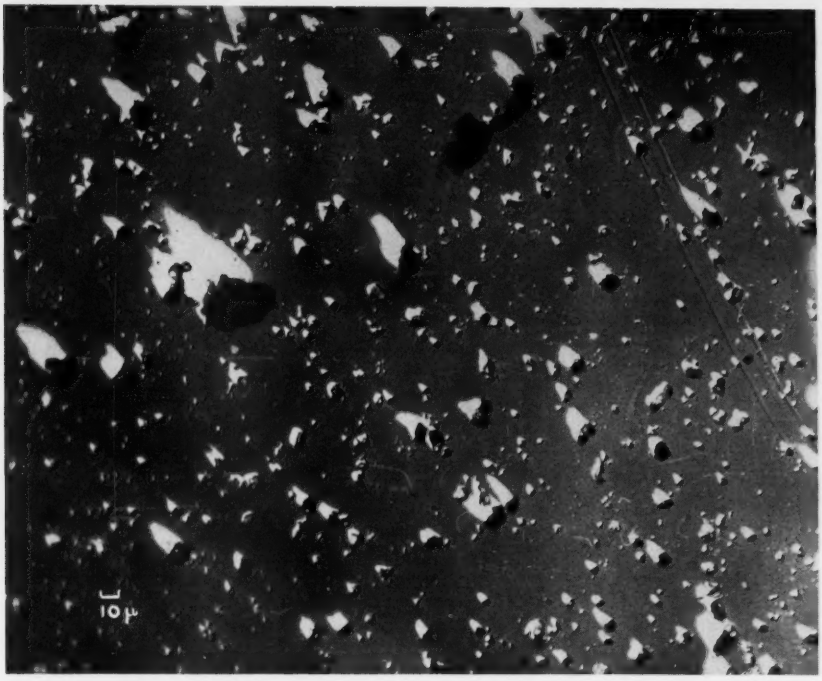


FIG. 1.—Photomicrograph of shadowed sample. 300×. Tangent of Shadow Angle 0.401.

source element is gold wire wrapped round the heating filament over as small an area as possible. The chamber is evacuated and the filament heated until the gold evaporates. The molecules of gold then disperse along straight lines from the point of origin and the slide is thus coated with a thin layer of gold, except in the areas shielded from the gold by the particles. These areas are called the shadows. The thickness of gold on the slide is negligible compared with the height of those particles which are large enough to be measured. The height of a particle can then be found from the angle of illumination and the length of the shadow. However, the height obtained depends on the shape assumed for the particle. This is discussed below. A relatively very thick deposit of gold is necessary for this work or the shadows will not be distinct enough. This is because the very powerful light, necessary for taking photographs through a microscope, obliterates the contrast between shadows and metal if the film is very thin. If the photographs were to be made with an electron microscope, using clays for example, a much thinner layer of gold would have to be deposited.

After shadowing, the slide is mounted on a petrological microscope using a high power lamp and convergent light. It was found that satisfactory photographs could be obtained using magnifications between 150× and 400×. The prints of these photographs could be enlarged to exactly 300× and 1000× respectively. Figure 1 shows a photograph of a sample from a Swedish varved sediment.

The photographs obtained are then put under a binocular microscope and magnified about ten times. A micrometer eyepiece allows the diameters and shadow lengths to be read off for each grain.

It will be found that not all the particles

are measurable, due either to aggregation, interference of shadows, or simply because they are too small to give a clear outline. In the author's work two hundred particles were counted from each sample.

The measurements taken are: the length of the shadow, L; the diameter of the grain parallel to the direction of maximum length, α; and the maximum and minimum diameters of the grain at right angles, α' and β. The α' diameter is measured only if α is not a maximum or a minimum diameter (fig. 2.)

The next step is the calculation of the actual height of each particle. The shadow angle is required for this. This is obtained from the average of about twenty angles calculated from the shadowed spheres. It is necessary to take into account the position of the spheres on the slide, since the evaporating metal is radiating from a small source. The angle of incidence of the rays therefore varies along both the length and the width of the slide. This variation must be allowed for by noting to the nearest millimetre the position of the photograph relative to the slide and calculating the glancing angle for that position, the angle being assumed constant over the area of one photograph.

The chief source of error in the method lies in the derivation of particle height from the shadow length, but before discussing this a brief analysis of the other sources of error will be made. The method of sampling is the first of these. As only a minute quantity of sediment can be used it is not possible to employ normal methods of random sampling. Particles above 20–30 microns cannot be transferred to the slide, since they settle too quickly to be drawn into the pipette. There is a lower size limit to measurement of about two microns with this method since anything smaller cannot be resolved by optical microscopes. For the finer grades a separate analysis could be made using electron photomicrographs.

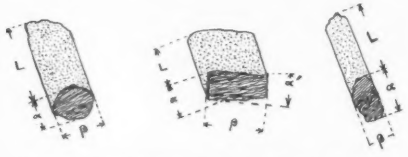


FIG. 2.—Measurements made on particles.

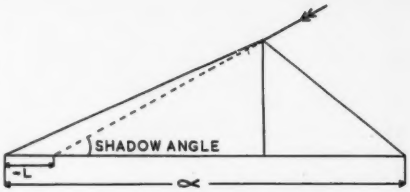


FIG. 3.—The shadowless particle.

The actual shadowing is a source of error since the shadowing metal does not come from a true point source and the shadows are thus inevitably not as sharp as they should be. This is due to the necessity of depositing a thick film of gold. A large piece of gold wire wound around the element will obviously evaporate from a length of the element wire and not one small area of it. There is thus great difficulty in obtaining sharply defined shadows.

The granules which were used to calculate the shadow angle may not have been perfect spheres but the amount of error introduced because of this is negligible.

DERIVATION OF PARTICLE HEIGHT

The last and greatest source of error is the irregularity of particle shape. The most obvious example is the shadowless particle (fig. 3). This occurs when the summit of a particle has the surface sloping away from it at a smaller angle than the shadow angle. Since the shadow angles are so small this condition is not very likely.

Particles are of widely varying shapes and for any one particle the calculated height clearly depends on what particle shape is initially assumed. This is the greatest difficulty of the shadow method and the relation between shadow length and particle height is treated in the following paragraphs.

Three possible shapes have been considered in order to find the maximum errors that can arise. They are:

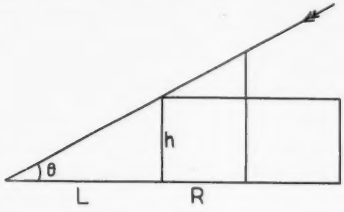
Case I. Sheer-sided particles, e.g., CYLINDERS.

Case II. Ellipsoidal particles, of which the sphere is a special case, e.g., ELLIPSOIDS.

Case III. Particles with an apex at their centre, e.g. CONES.

The formula for finding the height in each case is given in figure 4. The derivation of Formula II is given in the Appendix. The errors involved in using a particular formula on a particle of the wrong shape, knowing that the tangent of the shadow angle (θ) will vary from 0.2–0.5, are given in table 1, the maximum error in calculated height being given as a percentage of the height. It proved difficult to work out the errors involved in the case of an ellipsoid, so the errors are given for a sphere. The errors are larger for an ellipsoid than for a sphere but not greatly so. The error also varies with the actual dimensions of a particle, and to illustrate this two cases are quoted in table 1, one where the height equals the diameter α, and the other where it equals one tenth of the diameter α.

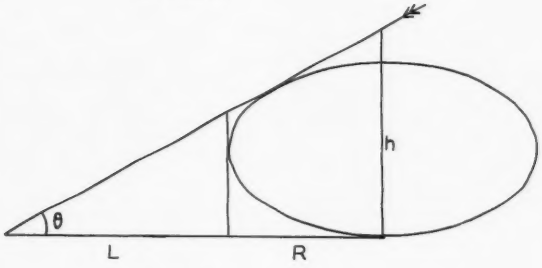
The ellipsoid formula was used for all particles where the measured diameter parallel to the shadow length, diameter α, was less than $4L \tan \theta$. For values of α greater than or equal to $4L \tan \theta$ the cylinder formula was used, thus assuming that all the particles whose ratio of α diameter to



CASE I

FORMULA I

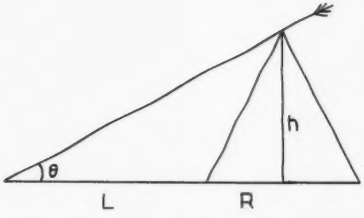
$h = L \tan \theta$



CASE II

FORMULA II

$$h = \frac{L \tan \theta (2R + L)}{R + L}$$



CASE III

FORMULA III

$h = (L + R) \tan \theta$

FIG. 4.—The three particle shapes considered.

TABLE 1

Error for	Formula used	Cylinder	Ellipsoid	Cone
Cylinder	When $h=\alpha$	None	20%	25%
	When $h=0.1\ \alpha$		70%	250%
Sphere		20%	None	5%
Cone	When $h=\alpha$	25%	6%	None
	When $h=0.1\ \alpha$	250%		

height is greater than 4 to 1 are sheer-sided. This is almost certainly true in the fresh sediment used by the author, since most, if not all, of the flat particles were mica flakes. In the case of particles that were in fact cylindrical but of greater ratio of diameter to height than 4:1 there will be an error. In the case of a cylindrical particle of ratio 3:1 this error will be 43% of the height.

To summarise: it has been shown that the calculated height of a particle is affected more by the assumed shape than by any of the other errors involved in the method. The best approximations to true particle shape are thought to be ellipsoids and cylinders, the cylinders being only the plate-like minerals. Therefore it is proposed to assume that all particles whose α diameter to height ratio is less than 4:1 are ellipsoids, and all others are cylinders. If any particles occur which are cylindrical and are treated as ellipsoids the errors could be as much as 50% of the calculated height. It is an error which increases the calculated height. In actual practise it would decrease the ratio of diameter to height from about 3:1 or 4:1 to about 2:1 or 3:1. This is the maximum possible error of the method.

PRESENTATION OF RESULTS

The obvious method of presenting three parameters is on a two-dimensional graph, but for ease of comparison of several samples, histograms are more convenient. For this reason the conversion of results into histograms of sphericity is recommended.

Wadell (1933) develops the expression for true sphericity from the unique isoperimetric property of a sphere. He states (p. 317):

"1. Solids of equal surface area and equal volume are the same shape.

"2. Solids of equal surface area but different volume have different shapes.

"3. Solids of equal volume but different surface area have different shapes."

Wadell uses the third statement to obtain the degree of true sphericity (ψ) where

$$\psi = s/S$$

S being the surface area of the particle, and s the surface area of a sphere of the same volume. However, he suggests (1935) the index $\phi = d_e/D_c$, where D_c = the diameter of the circumscribing circle of the projection of the greatest and median diameters, and d_e = the diameter of a circle equal in area to this projection, as being more practicable. He states that its value approaches very close to that of the degree of true sphericity. He also shows (1934) that ψ is a valid concept to use in work on the velocities of falling particles through water as it bears a mathematical relationship to the terminal uniform settling velocity.

Krumbein (1941) shows that Wadell's sphericity value ϕ (*not* Ψ as Krumbein says, p. 71) is equal to $\sqrt[3]{bc/a^2}$ for an ellipsoid—the "Intercept Sphericity." He also quotes four shape classes from Zingg (1935) and suggests that they are used in conjunction with ϕ, since Wadell's sphericity only relates surface area to volume. This being so, particles of different shape can have the same sphericity, e.g. a particle of ratios 2.5:1:1 has almost the same sphericity as one of 2.5:2.5:1, but clearly they are different shapes.

This Intercept Sphericity ϕ was in fact used by the author to analyse the results of the shape analyses, but since then Aschenbrenner (1956) has published a method which is preferable. Using a formula which

approximates the sphericity of a tetrakaidekahedron he has computed a chart of "Working Sphericity" ψ' (1956, fig. 4.) plotted on the co-ordinates b/a, c/b where $a>b>c$ are the three parameters of a particle. Also on the same chart are lines of equal Shape Factor (F) which, like Zingg's shape classes, is the ratio of the co-rodinates;

$$F=ac/b^2.$$

For each particle measured in a sample the ratios b/a and c/b are calculated, and the grain plotted as a point on the chart with these ratios as co-ordinates (Aschenbrenner, 1956, fig. 4). The number of grains within each of the ranges of sphericity and shape factor are counted and plotted as histograms, one for each characteristic. The fact that the two characteristics are independent of one another means that the two histograms are not necessarily related. However, it can be easily seen from the "spot chart" whether a peak on the sphericity histogram does or does not correspond to a peak on the shape factor histogram. The conclusions arrived at by a study of both histograms and the chart will then be a valid description of the sample. The only conceivable advance in the future would be the discovery of a method of combining these two characteristics to give the number of particles with an Equivalent Shape.

CONCLUSIONS

The method was checked by analysing a sample a second time, repeating all stages. Only one hundred particles were counted instead of two hundred, but the percentages within each sphericity range were substantially the same. The maximum difference was 6% of the whole and the standard deviation less than 3.5%, suggesting a quite remarkable degree of accuracy to the experiment.

The recommended method is, then, (1) to shadow a sample, dispersed on a slide; (2) take photomicrographs at 15× or 40×; (3) enlarge these to 300× or 1000×; (4) read off the two diameters and the shadow length; (5) calculate the height by the appropriate formula; (6) plot each grain as a point on a chart with axes b/a and c/b; and finally (7) construct histograms of sphericity and shape factor using Aschenbrenners' chart (1956).

In spite of limitations, this method of shape analysis would seem to be generally useful for silts and, using an electron microscope, clays. It is, of course, a slow method, but it can at least lay a claim to be a true expression of particle shape, which no two-dimensional method can.

ACKNOWLEDGEMENTS

This work forms part of a Ph.D. thesis submitted to Birmingham University, which provided a Research Scholarship to undertake the project. I am grateful to Dr. D. H. Griffiths and Dr. R. F. King for constant advice throughout the work and to Professor F. W. Shotton for the use of his Department. Mr. P. S. Garrett has been helpful in many ways and it was due to the helpfulness and patience of Mr. C. C. Newton of the Electron Physics Department that the shadowing technique was successfully evolved. Lastly I would like to thank Professor P. Allen of Reading University for kindly sponsoring the paper and also for his helpful criticism.

APPENDIX

Derivation of Formula for the Height of an Ellipse from Its Shadow Angle

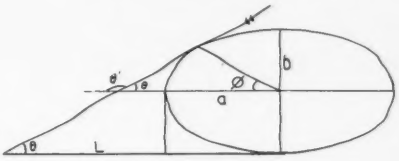


FIG. 5

Considering the case of an ellipse of semi-diameters a and b.

Tangent to the ellipse is

$$\frac{x}{a}\cos\phi+\frac{y}{b}\sin\phi=1. \qquad (1)$$

Gradient of tangent is

$$\tan\theta'=\frac{b}{a}\cot\phi \qquad (2)$$

$$\therefore \quad \cot^2\phi=\left(\frac{a}{b}\tan\theta'\right)^2. \qquad (3)$$

When

$$x=a+L, \qquad y=-b.$$

From (1)

$$\frac{a+L}{a}\cos\phi-\sin\phi=1$$

$$\therefore \quad \frac{a+L}{a}\cot\phi-1=\operatorname{cosec}\phi$$

From (2)

$$-\frac{a+L}{b}\tan\theta'-1=\operatorname{cosec}\phi$$

From (3)

$$-\frac{a+L}{b}\tan\theta'-1=\pm\sqrt{1+\frac{a^2}{b}\tan^2\theta'}$$

Square

$$1+2\frac{a+L}{b}\tan\theta'+\frac{(a+L)^2}{b}\tan^2\theta'$$

$$=1+\frac{a^2}{b^2}\tan^2\theta'.$$

Multiply by $b/\tan\theta'$

$$2(a+L)+\frac{1}{b}(a+L)^2-\frac{1}{b}(a^2\tan\theta')=0$$

$$2(a+L)+\frac{1}{b}(2aL+L^2)\tan\theta'=0$$

$$b=\frac{-L(2a+L)}{2(a+L)}\tan\theta'$$

$$b=\frac{L\tan\theta(2a+L)}{2(a+L)}$$

but $2b=h$

$$\therefore \quad h=\frac{L\tan\theta(2R+L)}{R+L}.$$

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FIGS. 1–4, SEPTEMBER, 1957

A NEW METHOD FOR OBTAINING SIGNIFICANT AVERAGE DIRECTIONAL MEASUREMENTS IN CROSS-STRATIFICATION STUDIES[1]

O. B. RAUP AND A. T. MIESCH
U. S. Geological Survey

ABSTRACT

A new method is described that can be used conveniently in the field to determine the number of measurements of cross-strata dip directions necessary to obtain a significant average direction for an area. The method commonly used in the past has involved a great deal of graphical computation in the field. The suggested method is based on the fact that the number of measurements needed is approximately proportional to their variation. The estimated standard deviation of the first 50 measurements is determined, and, by simple linear regression equation, the number of measurements needed is computed. This regression equation has a standard error of plus or minus 18 measurements. If 36 measurements are taken in addition to the number indicated by the equation, the geologist can be about 95 percent confident he has enough measurements to compute an average dip direction that will be comparable to that which would be obtained with the standard cumulative vector direction curve method. With the suggested method confidence intervals for the average directions can be obtained easily. A brief discussion of the application of Stein's two-stage sampling method to cross-stratification studies is also given. Stein's method is applicable to the case where an average direction must be within previously specified confidence limits.

INTRODUCTION

This paper describes a new field method that enables the geologist to determine, with *minimum effort, the* number of directional measurements of cross-stratification necessary to obtain a *significant average*[2] direction for any specific area.

Cross-stratification studies are now in progress as a part of a detailed stratigraphic study of rocks of Triassic age and associated formations of the Colorado Plateau by the U. S. Geological Survey on behalf of the Division of Raw Materials of the U. S. Atomic Energy Commission.

Churchill Eisenhart, National Bureau of Standards, reviewed an earlier draft of this paper and drew the writers' attention to the similarities and dissimilarities between the method here proposed and Stein's (1945) two-stage sampling method for achieving confidence intervals of fixed widths for the mean of a sampled population.

[1] Publication authorized by the Director, U. S. Geological Survey. Manuscript received April 29, 1956.

[2] "Significant average" as used in this report refers to an average which lies within previously specified confidence limits or an average for which confidence limits are known.

GEOLOGIC SIGNIFICANCE OF CROSS-STRATIFICATION

Cross-stratification is a common sedimentary structure in rocks of the Colorado Plateau. It has proved useful in reconstructing depositional environments, in stratigraphic correlation, and in the exploration for uranium and vanadium ore deposits of the Coloroado Plateau.

In reconstructing depositional environments, cross-stratification is interpreted to indicate directions of sediment transport and locations of source areas. The agent of deposition, velocity of current, and supplies of sediment are reflected in the degree of inclination and type of cross-strata.

Cross-stratification has proved to be a valuable tool in stratigraphic correlation. Similar types of cross-strata and direction of transportation suggest a similar environment of deposition. These similarities, supplemented with lithology and stratigraphic position, provide a basis for reliable correlation.

Exploration for ores of uranium and vanadium of the Colorado Plateau can be aided by an understanding of sedimentary structures (Stokes, 1954). Most of these deposits occur in channel-filling sandstones

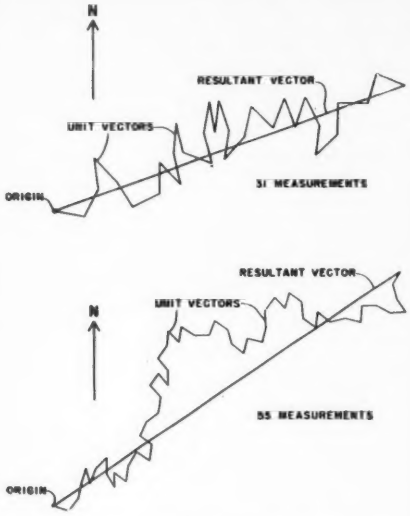

FIG. 1.—Typical resultant vector direction diagrams.

of the Shinarump Conglomerate and Chinle and Morrison Formations. Exploration is aided by determining the general trend of the channel from sedimentary structures at the outcrop. Successful application of cross-stratification in exploration for uranium and vanadium ore deposits of the Colorado Plateau is exemplified by the recent work of Stokes (1953, 1954) and Lowell (1955).

MEASUREMENT OF DIRECTIONAL COMPONENTS OF CROSS-STRATA

Because of variations in the directions of dip of cross-strata in any given area it is necessary for purposes of interpretation to determine the average direction. The average direction of a set of measurements can be determined most easily by a graphical method illustrated in figure 1. Using this method, directions of dip of cross-strata are plotted on vector diagrams. Each measurement is represented by a vector of unit length. A straight line that connects the origin with the end of the last vector is the *resultant vector*. The resultant vector is an estimate of the average dip direction of cross-strata in the area if the initial measurements were taken at randomized intervals over the area. One measurement is made per set of cross-strata, and the location of measurements is controlled by the local geology. The number of measurements necessary to obtain a significant estimate of the average direction is approximately proportional to the variation of cross-strata dip directions in the area. Depending on the geology, this variation may be a function of the size of the area. Three methods of determining the number of measurements necessary are given below.

Standard Method

A method described by Reiche (1938) for determining the number of cross-strata dip direction measurements necessary to obtain a significant average direction in an area has become more or less standard among geologists. With Reiche's method, successive average directional measurements (each taken with 10 additional measurements for this report) when plotted on a graph define a "cumulative vector direction curve" (fig. 2). The curve normally is highly irregular near the origin, where it is based on few measurements, but toward its other extremity it becomes essentially horizontal. Reiche (1938) refers to the number of measurements at which the curve approaches and beyond which it remains within a 5 degree range as the "flatness point." The curve can be extended any number of measurements beyond the flatness point, depending on the degree of verification desired. For this report the flatness point is considered the point beyond which the cumulative curve remains within a 5 degree range for 2 additional groups of 10 measurements.

Confidence limits for average directional measurements can be determined by the commonly used equation,

$$h=\frac{ts}{\sqrt{n}}$$

where h is the half-width of the confidence interval, s is the standard deviation, n is the number of measurements, and t is taken from a table of t for a given probability (see, for example, Fisher and Yates, 1953, p. 40). Average directions computed on the number of measurements indicated by the flatness point, therefore, have confidence

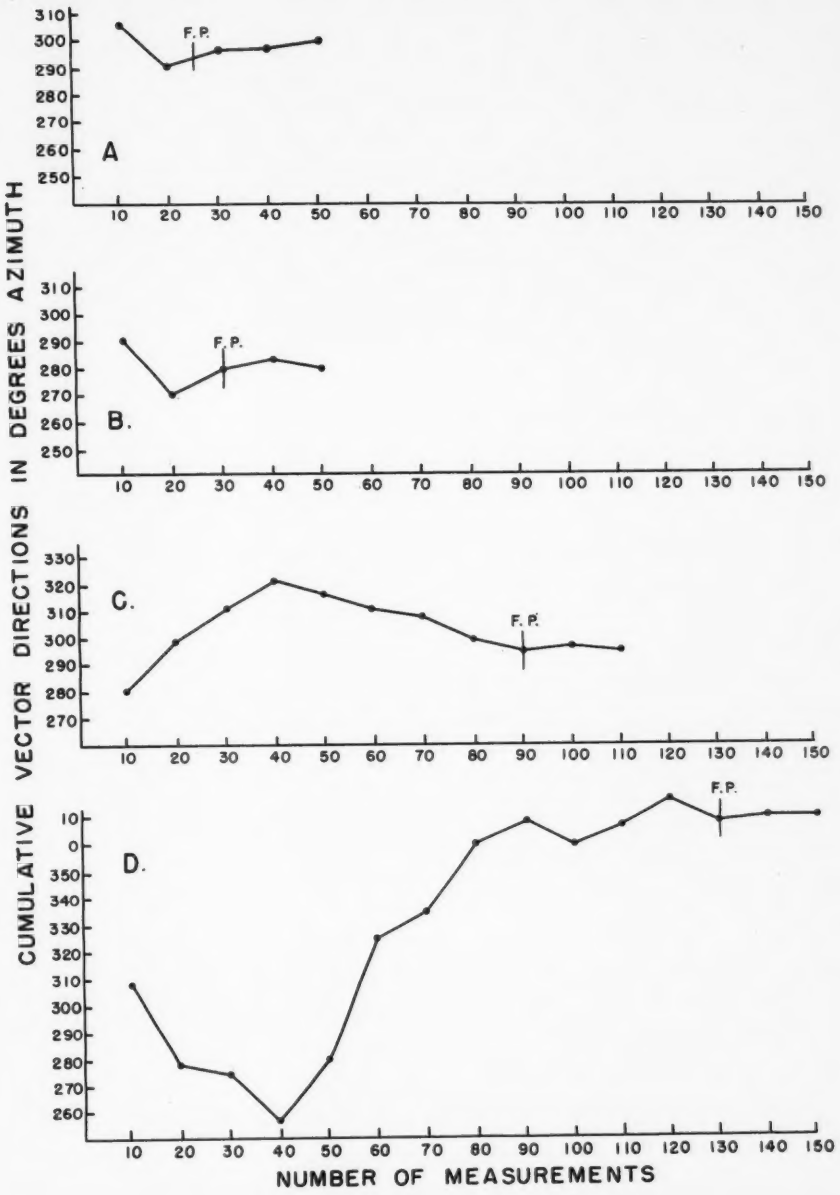

FIG. 2.—Cumulative vector direction curves.

limits proportional to the flatness point and the variation of the measurements. The 0.95 confidence intervals are almost always larger than $\pm 5°$, the range to which the cumulative vector direction curve is confined beyond the flatness point.

The flatness point method of determining the number of directional measurements

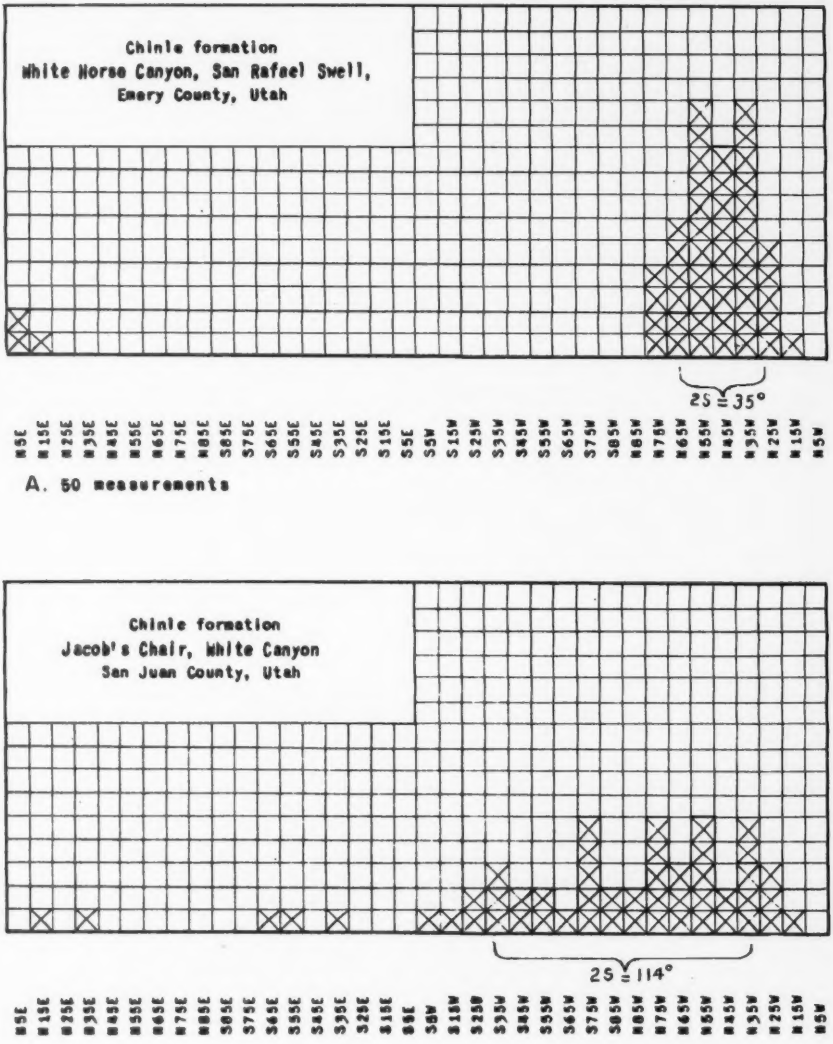
Chinle formation
White Horse Canyon, San Rafael Swell,
Emery County, Utah
25 = 35°
N5E N15E N25E N35E N45E N55E N65E N75E N85E S85E S75E S65E S55E S45E S35E S25E S15E S5E S5W S15W S25W S35W S45W S55W S65W S75W S85W N85W N75W N65W N55W N45W N35W N25W N15W N5W
A. 50 measurements
Chinle formation
Jacob's Chair, White Canyon
San Juan County, Utah
25 = 114°
N5E N15E N25E N35E N45E N55E N65E N75E N85E S85E S75E S65E S55E S45E S35E S25E S15E S5E S5W S15W S25W S35W S45W S55W S65W S75W S85W N85W N75W N65W N55W N45W N35W N25W N15W N5W
B. 50 measurements

FIG. 3.—Frequency distribution histograms.

necessary in an area requires that the geologist calculate a relatively large number of average directions in the field and that he plot these as cumulative curves to determine the flatness point. The labor of computation involved in this procedure makes it undesirable and an alternative method is suggested below.

Suggested Method

The suggested method for determining the number of directional measurements necessary in an area to obtain a significant average direction is based on the fact that the flatness point of a cumulative vector direction curve is roughly proportional to the variation in dip direction of the cross-strata in the area. An estimate of the variation in dip direction can be made easily in the field. Fifty measurements of cross-strata dip directions are taken and plotted on histograms similar to those shown in figure 3. Each histogram, in reality, is plotted on a circular base and if cross-strata dip directions in an area are purely random it is possible that the histogram will have approximately equal frequencies in each class interval all around its base. Most often, however, the dip directions have an approximate *normal* (or "circular normal") distribution and the measurements will cluster at some part of the base, forming a distinct mode. For purposes of mathematical simplicity the distribution will be treated as normal, which, for the observed distributions, seems to be a satisfactory approximation. When 8, or 16%, of the 50 measurements are subtracted from each tail of the histogram (measurements are subtracted from classes farthest from the mode), the range in degrees of the central 34 measurements, or 68%, can be determined approximately. Because only part of the number of observations are commonly subtracted from certain class intervals, some interpolation is necessary. For example, if 4 out of 5 measurements in a class interval are to be subtracted, the range of the central 68% extends about one-fifth of the width of the class interval, or 2°, into the class interval (fig. 3a). The range determined by this procedure approximates the range covered by the measurements within 1 estimated standard deviation of the mean and is referred to here as 2 estimated standard deviations ($2s$).

The linear correlation coefficient between the ($2s$) values determined by this method and the flatness points of 35 series of measurements in various areas and stratigraphic units on the Colorado Plateau has been found to be ± 0.86 (fig. 4). The regression equation expressing the variation of the flatness point with ($2s$) is: F.P. $= 0.65$ ($2s$) $- 8.7$, where F.P. is flatness point or number of directional measurements necessary in an area to obtain a significant average direction. The equation has a standard error of plus or minus 18 measurements. If 36 measurements are taken in addition to those required by the equation to reach the flatness point, one can be more than 95 percent confident that, had a cumulative vector direction curve been plotted, the flatness point would have been reached. The resultant vector of these measurements can then be determined in the office where more careful graphical plotting can be done.

Two standard errors of estimate (36 measurements) are used here to insure attainment of the 0.95 confidence limit. The 0.95 confidence limit ($d_{0.95}$) is determined more precisely from an expression given by Eisenhart (1939, p. 180):

$$d_{0.95} = t_{0.95} S_{y/x} \left\{ 1 + \frac{1}{n} + \frac{(x-\bar{x})^2}{\Sigma (x-\bar{x})^2} \right\}^{1/2},$$

which allows for the experimental uncertainty of the fitted regression line. Using this equation the confidence limit increases with increasing value of the term $(x-\bar{x})^2$, and at x, or ($2s$) $= 210$, or ($2s$) $= 0$, 33 additional measurements are required. This is in substantial agreement with the 36 measurements specified by the use of two standard errors.

The standard error of estimate of the regression equation as used above is defined by the equation:

$$S_{y/x} = \sigma y \sqrt{1 - r^2},$$

where σy is the standard deviation of the flatness point estimates and r is the linear correlation coefficient between values of ($2s$) and values of flatness point.

In about 95 percent of the cases either the first 50 measurements or the flatness point determined from the regression equation

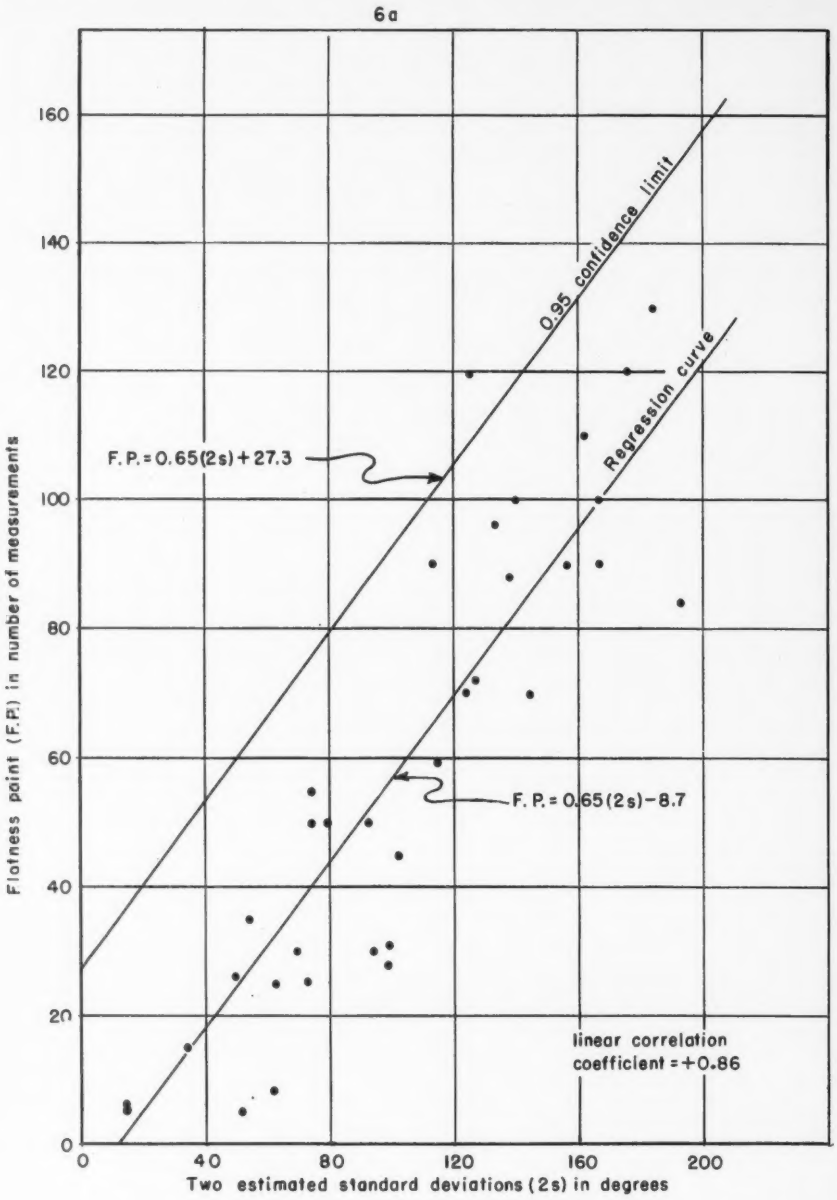

FIG. 4.—Estimated (2*s*) and flatness point correlation graph.

TABLE 1.—*Number of additional measurements required (beyond the first 50) for values of (2s) up to 200°*

(2s) degrees	Additional measurements required	0.95 confidence limits of resultant vector (degrees)
≦35	0	± 5
36–37	1	± 5
38	2	± 5
39–40	3	± 5
41	4	± 6
42–43	5	± 6
44	6	± 6
45–46	7	± 6
47–48	8	± 6
49	9	± 6
50–51	10	± 6±7
52	11	± 7
53–54	12	± 7
55	13	± 7
56–57	14	± 7
58	15	± 7
59–60	16	± 7
61	17	± 7
62–63	18	± 8
64	19	± 8
65–66	20	± 8
67–68	21	± 8
69	22	± 8
70–71	23	± 8
72	24	± 8
73–74	25	± 8±9
75	26	± 8
76–77	27	± 9
78	28	± 9
79–80	29	± 9
81	30	± 9
82–83	31	± 9
84	32	± 9
85–86	33	± 9
87–88	34	± 9±10
89	35	±10
90–91	36	±10
92	37	±10
93–94	38	±10
95	39	±10
96–97	40	±10
98	41	±10
99–100	42	±10
101	43	±10
102–103	44	±11
104	45	±11
105–106	46	±11
107–108	47	±11
109	48	±11
110–111	49	±11
112	50	±11
113–114	51	±11
115	52	±11
116–117	53	±11±12
118	54	±12
119–120	55	±12
121	56	±12
122–123	57	±12
124	58	±12
125–126	59	±12
127–128	60	±12
129	61	±12
130–131	62	±12
132	63	±12
133–134	64	±12±13
135	65	±13
136–137	66	±13
138	67	±13
139–140	68	±13
141	69	±13
142–143	70	±13
144	71	±13
145–146	72	±13
147–148	73	±13
149	74	±13
150–151	75	±13±14
152	76	±14
153–154	77	±14
155	78	±14
156–157	79	±14
158	80	±14
159–160	81	±14
161	82	±14
162–163	83	±14
164	84	±14
165–166	85	±14
167–168	86	±14
169	87	±14
170–171	88	±14±15
172	89	±15
173–174	90	±15
175	91	±15
176–177	92	±15
178	93	±15
179–180	94	±15
181	95	±15
182–183	96	±15
184	97	±15
185–186	98	±15
187–188	99	±15
189	100	±15
190–191	101	±15±16
192	102	±16
193–194	103	±16
195	104	±16
196–197	105	±16
198	106	±16
199–200	107	±16

plus 2 standard errors (36 measurements), will be greater than the flatness point determined by the standard method. Where $(2s)$ is less than 36°, values of (F.P.) plus 2 standard errors will be less than 50 and no additional measurements are necessary. Table 1 gives the additional measurements required for values of $(2s)$ from 36° to 200°.

Table 1 also gives the 0.95 confidence limits for resultant vectors computed for various numbers of measurements with corresponding values of $(2s)$. It will be noted that the confidence limits increase with increasing values of $(2s)$ and increasing numbers of measurements. That is, when $(2s)$ for the first 50 measurements is 36°, one additional measurement is taken and the resultant vector is, with 0.95 probability, within ±5° of the true average direction. When $(2s)$ for the first 50 measurements is 200°, 107 additional measurements are taken and the resultant vector is, with 0.95 probability, within ±16° of the true average direction. It may be necessary in some cases to reduce the confidence limit of the second resultant vector to ±5° as was obtained with the first set of measurements. To do so, however, it will be necessary to take 1566 additional measurements rather than 107. The basis of this estimate is given in the following section.

Stein's Two-Stage Sampling Method

A method devised by Stein (Stein, 1945; Cochran, 1953, p. 59–61) can be applied to the problem of determining the number of cross-strata dip directions measurements necessary to compute an average direction within previously specified 0.95 confidence limits. The standard deviation of the first 50 measurements may be inserted in the equation,

$$n=\frac{4.04s^2}{d^2}$$

where d is the desired half-width of the previously specified confidence interval. For example, with a standard deviation of 20° it is necessary to take 65 measurements to compute an average direction within ±5° of the true average direction,

$$n=\frac{4.04s^2}{d^2}=\frac{4.04\times(20)^2}{(5)^2}=64.6$$

In this case 15 additional measurements should be taken. Using the same equation it can be seen that when the standard deviation of the first 50 measurements is 100° $(2s=200)$ 1566 additional measurements must be taken to reduce d to 5°.

SUMMARY

Histograms for the estimation of $(2s)$ for the initial 50 measurements can easily be constructed on the cross-section paper of a standard field notebook. In using the suggested method it is not required that the geologist compute resultant vectors in the field, nor is it necessary for him to construct cumulative vector direction curves. He can refer to table 1 which gives him the number of additional measurements required. After taking the indicated number of additional measurements he can be about 95 percent confident that, had he constructed a cumulative vector direction curve, he would have reached the flatness point. The resultant vector of the total measurements can be determined at the geologist's convenience. The method is not suggested as a way to lessen the number of measurements taken for it will certainly require a few more, but, rather it is suggested as a way to eliminate graphical computation in the field, an ever cumbersome task for the field geologist.

The 0.95 confidence interval of the resultant vector can be read from table 1 for values of $(2s)$ for the first 50 measurements and the number of additional measurements taken. The geologist can be 95 percent confident that the true average dip direction lies within these limits.

As shown by table 1 the confidence limits increase with values of $(2s)$ and the number of measurements. A number of resultant vectors determined by this method when plotted on a map, therefore, would not all have the same reliability; their 0.95 confidence intervals will vary up to ±16° when $(2s)$ for the first 50 measurements equals 200°. An error of 16° is not too large to prevent valid geologic interpretation, especially when it is considered that geologic interpretation is rarely based on a single vector but on a number of vectors plotted on a map. The number of measurements necessary to reduce the confidence interval to ±5° may be determined with Stein's equa-

tion, though when ($2s$) of the first 50 measurements is large the number of measurements necessary becomes impractical, if not impossible, to collect.

In the case where a resultant vector or average direction must be determined within previously specified confidence limits Stein's equation may be applied by using the standard deviation of the first 50 measurements computed with the conventional formula for standard deviation.

The relationship between flatness point (F.P.) and variation of directional measurements ($2s$) was observed within stratigraphic units on the Colorado Plateau, but it probably can be extended to other areas. However, the confidence intervals given in Table 1 will be applicable in all cases to corresponding values of ($2s$) and number of measurements. Stein's two-stage sampling method is completely independent of the flatness point concept.

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 27, NO. 3, PP. 322–332
FIGS. 1–2, SEPTEMBER, 1957

THE NATURE AND ORIGIN OF ARGENTINE LOESS[1]

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ABSTRACT

Though the loess deposits of Argentina (Pampean Formation) are geologically well known, their granulometric and mineralogical composition has hitherto not been analyzed. The data obtained from a study of more than fifty samples are here summarized. It is shown that Argentine loess is similar in field appearance and texture to the North American and European loess. However, the mineralogical composition is completely different, especially in the coarse fractions where an assemblage of volcanic minerals is predominant. The presence of abundant montmorillonite and glass shards in the fine fraction is a point of resemblance between the Argentine and the American sediments. All this material, doubtless of volcanic-pyroclastic origin, has been transported by winds to the place of deposition.

INTRODUCTION

In a recent paper, Swineford and Frye (1955) have made a substantial contribution to our knowledge of the petrography of loess by comparing samples from western Europe with those of Kansas. As is well known, there are two other regions in the world where loess and loess-like deposits cover vast areas: China and Argentina. The Argentine deposits, which are the most important in the southern hemisphere, have been known since the middle of last century from descriptions by D'Orbigny (1842), Darwin (1846), Bravard (1857), Ameghino (1881), and many others. However, very few petrographic data have been available, as the previous work has been mainly stratigraphic, with the exception of the study of some isolated samples by Wright and Fenner (1912) and some partial observations by European authors (Meigen and Werling, 1915; Principi, 1915). In order to fill this gap in our knowledge, a detailed investigation on typical samples were carried out by the author and collaborators (M.E.C. de Di Lorenzo and J. Remiro) at the Instituto Nacional de Investigación de las Ciencias Naturales y Museo of Buenos Aires. The results obtained show that the Argentine loess is entirely different in mineralogical composition from the European and North American loess, and probably from the Chinese as well (Viglino, 1901; Barbour, 1927).

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CHARACTERISTICS AND STRATIGRAPHY OF ARGENTINE LOESS

The vast distribution of loess deposits in Argentina is shown in figure 1. These Quaternary sediments are grouped under the collective name of Pampean Formation or, in short, Pampean, a name first used by D'Orbigny (*argile pampéene;* 1842) and later by Darwin (*Pampean Mud* or *Pampean Formation;* 1846). The average thickness of the formation is 90 feet; the maximum registered is 210 feet. The Pampean sediments are light yellow or brown in color, sometimes with a reddish or gray tinge; they are devoid of stratification, stand for a long time in vertical walls, possess tubes and rods of calcium carbonate formed around plant roots, contain remains of vertebrate fossils and have all the other characteristics typical of loess deposits the world over. It is not strange then that two European geologists (Heusser and Claraz, 1866) early recognized their resemblance to similar loess deposits of the Rhine valley.

Not all the Pampean sediment is really a loess, although the greater part of it has the appearance of such. It is now generally accepted that reworked or secondary loess is nearly as abundant, or even more abundant, than primary loess. It is extremely difficult to distinguish between these two, and in many cases almost impossible. Argentine geologists have been using, for some 30 years, the adjective *loessoid* to denote material with appearance of loess, regardless of origin. Frenguelli (1925) analyzed all the geological characteristics that may be used, at least in Argentina, to dis-

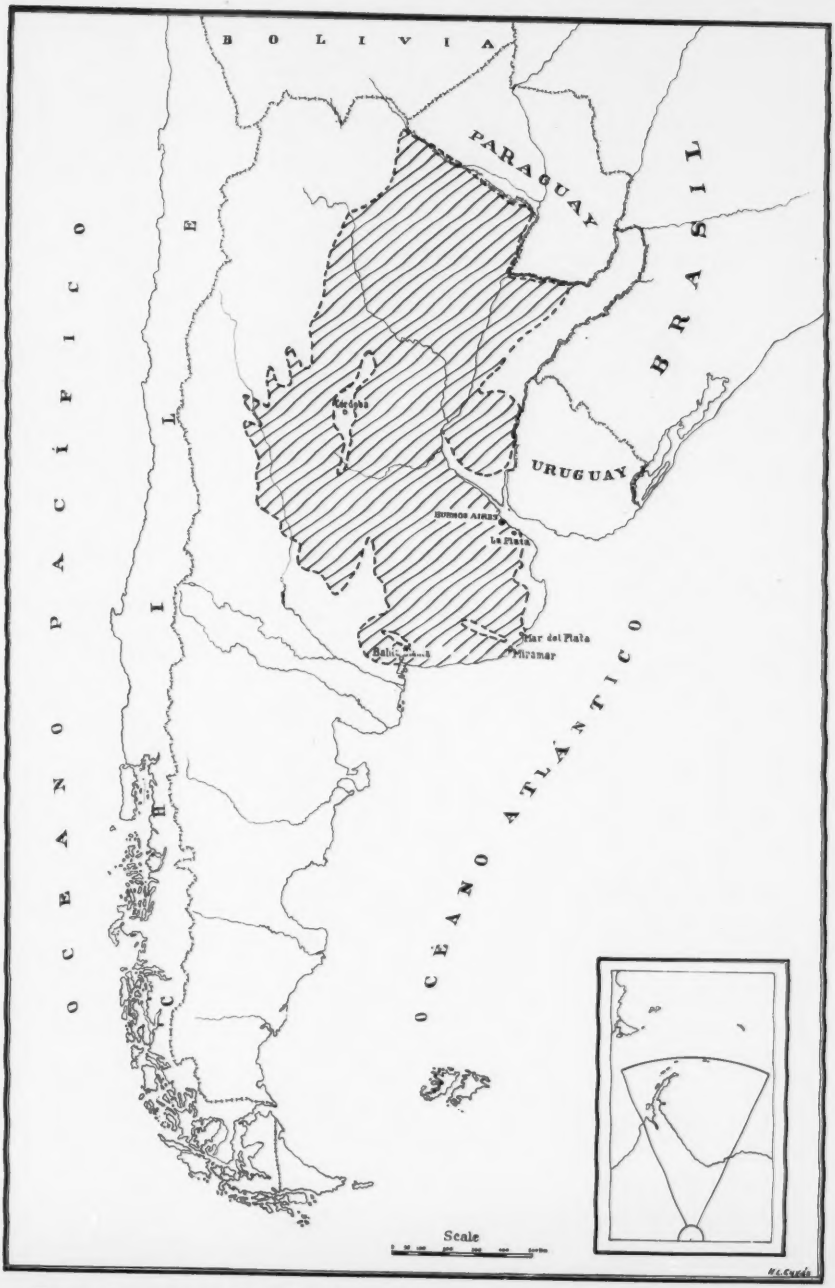

FIG. 1.—Approximate distribution of loessoid sediments (Pampean Formation) in Argentina.

tinguish primary from secondary loess. His contribution is of great value, even nowadays, though further systematic investigation of these deposits is necessary to elucidate this debated question (Smith, 1949).

The stratigraphy of the Pampean deposits was first established by Ameghino (1881). In table 1 the main stratigraphic units are listed, with their approximate equivalents in North American stratigraphy. The names introduced by Kraglievich (1952) are locally applied to the formations which he recognized in the coastal cliffs between Mar del Plata and Miramar, where the best and most complete section of the Pampean Formation is exposed. It has to be borne in mind that no close-fitting equivalence between Argentine and North American (or for this matter European) stratigraphic units is possible, as glaciations in Argentina were either small or nonexistent.

SAMPLING

Most of the samples were collected by J. L. Kraglievich from the Mar del Plata-Miramar cliffs. Other samples were collected at the same place by the author in February 1954. The sampling was done on surface exposures, where the stratigraphic relationships were clear.

C. Cortelezzi collected a certain number of samples from La Plata. These were obtained from excavations for the building of new water mains. No auger samples were studied.

The sampling covers practically all the Pampean Formation, i.e., from the lowest Pleistocene up to the present. The results obtained were also checked with samples collected by the author at different points of Buenos Aires Province.

LABORATORY PROCEDURE

Particle-size analysis.—Mechanical analyses were made by a method different from the one used by Swineford and Frye (1951; 1955). The samples were first treated with cold diluted (N/10) hydrochloric acid until effervescence ceased. The dispersion of the filtered material was made with distilled water, to which lithium carbonate in 0.2% solution was added until the pH of the suspension was 8.3. The choice of the peptizer was made after numerous trials, since all the common substances used in mechanical analyses (ammonia, sodium carbonate, sodium silicate, sodium oxalate, etc.) failed to produce a good dispersion of

TABLE 1.—*Stratigraphic divisions of the Pampean Formation of Eastern Argentina*

EASTERN ARGENTINA (Frenguelli, 1950)		MAR DEL PLATA-MIRAMAR AREA (Kraglievich, 1952; Groeber, 1952)	
Cordobense Platense	Epipluviar		Post-Wisconsin (Post-würm)
Querandinense Lujanense	Fourth Pluviar	Lobería	Wisconsin (Würm)
Bonaerense	Third Interpluviar	Cobo	Peorian (Riss-Würm)
	Third Pluviar	Santa Isabel	Illinoian (Riss)
Ensenadense	Second Interpluviar	Arroyo Seco	Yarmouth (Mindel-Riss)
	Second Pluviar	Miramar	Kansan (Mindel)
Chapalmalense	First Interpluviar	San Andrés	Aftonian (Günz-Mindel)
	First Pluviar	Vorohué	Nebraskan (Günz)
		Barranca Lobos	Danube Glaciation
Pliocene		Chapadmalal	Upper Pliocene

the samples. Therefore, our analyses are not entirely comparable with those of Swineford and Frye; however, it is thought that the differences in results ought not to be very great. The agitation of the samples was done by hand during a period of 24 hours; the use of laboratory shakers was precluded as in many cases they produced a strong coagulation. Some of the samples were so compact that they had to be disaggregated with a rubber pestle in a mortar prior to the chemical treatment. All the mechanical analyses were made by the pipette method.

Microscopic study.—The sand and silt fractions were examined with a petrographic microscope. The separation of heavy minerals was made with bromoform. The clay fraction was also observed under the microscope, after the pipette samples were evaporated on a glass slide. The mineralogical composition was determined by counting.

Determination of the clay fraction.—Besides the microscopic study, X-ray diffraction patterns of the less-than-1μ fraction of some samples were obtained with a "Debyeflex" (Rich, Seifert & Co.) unit. The patterns were interpreted by Engineer M. Butschowskyi, Instituto Nacional de Investigación de las Ciencias Naturales.

TEXTURE

Of a total of more than 50 analyzed samples, 12 were selected to illustrate the textural composition; the remaining unillustrated ones do not differ in appreciable degree from the ones here selected. With the exception of two samples from La Plata (134 and 20, analyzed by C. Cortelezzi), all the samples selected are from the cliffs between Mar del Plata and Miramar. They range from the lowest Pleistocene to the Recent.

The results of the mechanical analyses show a similarity in texture between the Argentine loess and the North American and European samples, as can be seen by comparing the values of table 2 and the cumulative curves of figure 2 with the data published by Swineford and Frye (1955).

In Argentine loessoid samples the value of Mϕ varies between 4.72 and 6.04, i.e., within the same range as in European and American loesses, or even smaller. Equally

TABLE 2.—*Mechanical analyses of loessoid samples from Eastern Argentina*

No.	Formation	Size distribution in phi units (percent by weight)											Mϕ	$\sigma\phi$	Mdϕ	QDϕ	Skϕ
		0–1	1–2	2–3	3–4	4–4.5	4.5–5	5–6	6–7	7–8	8–9	9–10	10				
62	Lobería		1.0	2.0	32.2	17.3	12.3	9.1	6.9	6.7	3.5	3.5	5.5	5.12	2.10	4.41	1.19	0.585
15	Santa Isabel			0.5	43.3	35.7	6.2	0.6	0.6	0.2	3.8	5.5	3.6	4.72	1.93	4.07	0.32	1.028
58	Miramar			2.9	34.0	22.5	12.2	10.2	7.2	7.9	1.7	0.6	0.8	4.73	1.49	4.26	0.77	0.495
8	Arroyo Seco			1.0	48.6	24.1	7.1	2.7	1.6	0.3	3.1	5.0	6.5	4.83	2.15	4.13	0.40	0.874
11	Barranca Lobos			0.2	41.1	19.6	8.1	5.5	3.3	2.8	4.9	6.6	7.9	5.29	2.25	4.18	1.19	0.600
50	Cobo			2.7	47.4	15.0	11.2	1.9	2.9	3.6	0.7	4.6	10.0	5.02	2.37	4.00	0.69	0.728
2	Vorohué			0.8	25.5	35.7	6.1	6.8	1.4	3.4	3.8	8.5	8.0	5.48	2.32	4.28	1.06	0.577
24	San Andrés			0.5	42.3	8.9	12.5	4.8	0.6	0.6	5.0	7.7	11.1	5.40	2.55	4.43	1.18	0.558
60	Vorohué	9.9	2.0	1.1	16.3	24.5	10.1	11.1	3.9	2.1	5.3	8.6	5.1	5.06	2.64	4.40	1.03	0.193
38	Chapadmalal				16.8	29.1	10.3	10.7	3.2	1.8	6.2	7.9	14.0	6.04	2.49	4.70	2.21	0.404
134	Ensenadense				23.5	12.5	11.3	29.7	2.5	6.5	0.5	0.5	13.0	5.43	2.19	5.10	0.92	0.774
20	Ensenadense				35.9	12.5	15.5	10.5	8.3	5.5	2.5	1.2	8.1	5.24	2.10	4.55	1.15	0.674

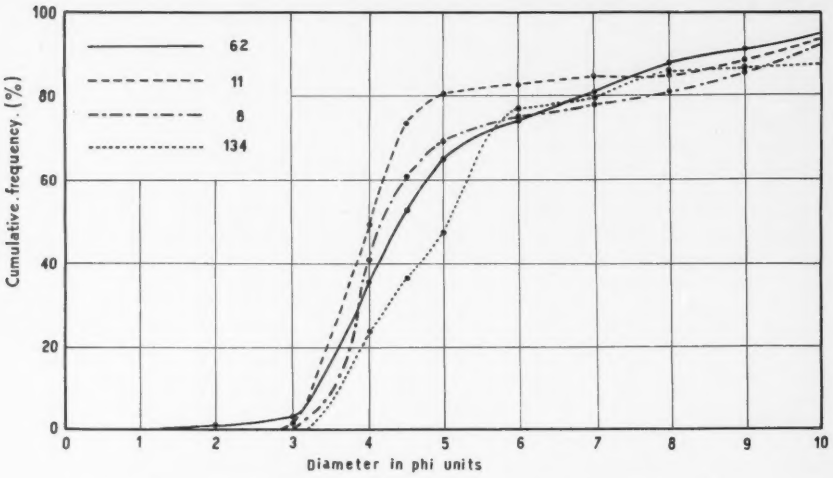

FIG. 2.—Selected cumulative curves for loess samples from eastern Argentina.

similar are the values for Mdϕ, though it must be pointed out that in Argentine samples there is a tendency for Mdϕ to be smaller (4.00 to 4.40 are the commonest values, as against values larger than 4.40 and up to 6.55 in samples from the Northern hemisphere). The close agreement in Mϕ and Mdϕ is even more remarkable when compared with Peorian loess from Nebraska and Kansas (Swineford and Frye, 1951).

The phi quartile deviation (QDϕ) is also comparable, though it tends to be slightly higher in Argentine samples; it ranges from 0.32 to 2.21. No significant differences exist in the values of phi skewness (Skϕ) and phi standard deviation ($\sigma\phi$), though the latter is generally larger in Argentine samples.

COMPOSITION

Sand Fraction

The composition of the "light minerals" fraction is fairly constant, the main constituents being plagioclases, quartz, orthoclase, volcanic glass, altered feldspars, fragments of volcanic rocks and organic opal.

Plagioclases are the most abundant minerals, in all the samples, varying in proportion from as much as 65% down to 20%; normally, however, they make up from one third to one half of the light minerals. They appear as irregular or cleavage fragments, somewhat rounded and usually showing polysynthetic twinning, Carlsbad-albite and albite being the most frequent laws. They are always remarkably fresh, a fact that did not escape recognition by Wright and Fenner (1912). U-stage determinations showed that the most common species is acid labradorite, followed by andesine, oligoclase (rare), and a variable quantity of albite (between 1 and 10%). Many of the plagioclase grains are zoned, and the presence of solid inclusions and "negative crystals" is to be mentioned.

Quartz is not very abundant; the maximum found was 30%, but more frequently it does not exceed 20% and can even be as low as 2%. The grains are subangular to subrounded, some with liquid inclusions which occasionally are arranged in rows. Grains without inclusions are very common, but those with solid ones are extremely rare. Wavy extinction is practically absent, as are secondary growths and ferric coatings.

Orthoclase appears in rounded grains, usually altered to some clay mineral (kaolin, allophane?). Microcline is extremely scarce, but when found is always unaltered. On the other hand, there is in all the samples a

variable proportion of grains so greatly altered that optical determination is extremely difficult; the greater part, however, seem to be derived from some potash feldspar.

Volcanic glass is always present in the form of angular shards; its percentage varies between 1 and 25% (not counting the thin intercalations of volcanic ashes found in most sections of Pampean loess). It is a colorless variety (rarely light brown), usually with flow structure and with refractive index between 1.49 and 1.51. These acid shards are always fresh and show no signs of alteration. Basic green or brown glass is very rare and is confined to one single level within the whole thickness of the Pampean Formation.

Rounded fragments of volcanic rocks, especially groundmass, are very characteristic in the Argentine loessoid sediments. Some of them are basaltic or andesitic in nature, while others are rhyolitic. Recognition is sometimes difficult because of alteration and impregnation with iron oxides and hydroxides. Their proportion varies between 1 and 25%.

A very typical and frequent constituent of Pampean loess is organic opal, found as minute saw-edged tablets which are really siliceous epidermic cells of grasses. When their form is lost by comminution, the small particles resemble irregular fragments of volcanic glass and only by the determination of their refractive index can they be distinguished (Teruggi, 1955). The percentage of opal is normally low (less than 1%), but in some cases it is as high as 5%.

Some other constituents are occasionally found, such as chalcedony, chert, sanidine, gypsum, fragments of argillaceous rocks, etc., but never more than one or two grains per sample.

The "heavy minerals" make up between 0.6 and 6% of the total sand fraction; in most samples, however, they vary between 0.7 and 1.5%. The principal components are iron ores, amphiboles, and pyroxenes, with a minor amount of other accesory minerals.

Opaque iron ores constitute the greater part of the heavy minerals. Magnetite is by far the predominant species and appears as well-rounded grains which under reflected light show signs of alteration into hematite; hematite is also found as independent grains of similar form and size. The magnetite is a titaniferous variety, as can readily be shown by chemical analysis; its presence in the Pampean sediments has been known since the middle of the last century (Bravard, 1857; Heusser and Claraz, 1866), when the names of "fer titané" and iserite were given to this species. Ilmenite is also common, but its percentage is small in comparison with that of magnetite. Limonite and leucoxene make up a small part of the opaque ores; pyrite is extremely rare and found in few samples. Many of the opaque grains, especially those of hematite and limonite, could very well be fragments of basic igneous rocks deeply penetrated by these secondary iron products.

The amphiboles are represented by a green or brown green hornblende, which appears as subrounded prismatic grains; it is always very fresh, in spite of showing signs of having endured long transportation. After the ores, it is the most abundant heavy mineral; it is always accompanied by smaller amounts of dark brown lamprobolite. Fibrous fragments of actinolite are occasionally seen.

The pyroxenes, nearly as abundant as the amphiboles, are hypersthene and augite. Hypersthene is found in long or short prisms, which sometimes are perfectly rounded and other times are fractured across or have the hacksaw terminations typical of intrastratal solution. Oriented rutile or opaque inclusions are frequent; pleochroism is very weak, from colorless to extremely pale brown. Augite appears as rounded grains, ovoidal in shape and dark green in color; in most samples a pale green diopsidic variety is also present in irregular fragments. Both clino- and orthopyroxenes are extremely fresh and clear, though they may have suffered strong attrition. Colorless enstatite is confined to samples from the northern part of Buenos Aires Province; its percentage is always very low.

The remaining minerals are very scarce, with the exception of zoisite and epidote, which in certain levels of the Pampean Formation may be common; they usually appear as rounded grains and prisms. Garnet is found in practically all the samples, but in small quantities; it is a colorless

or light red variety, which sometimes shows signs of attrition. Micas, both muscovite and biotite, may be found as occasional flakes; the same happens with the chlorites. Tourmaline, zircon, titanite, and rutile are occasionally present, but rarely exceed one or two grains in a sample. The systematic absence of apatite is noteworthy.

Silt Fraction

The composition of the silt fraction is, on the whole, similar to that of the sand fraction, but with one important difference: in all the samples there is a sharp increase in the contents of fresh and altered volcanic glass shards.

Coarse silt (0.062–0.031 mm): The percentage of glass shards varies between 15 and 60%, amounts of 30 and 40% being the most common. The other main minerals are plagioclase, quartz, and orthoclase, in the order mentioned. In some samples, montmorillonite replaces fresh volcanic glass; it appears under the form of altered and more or less fragmented shards. Gypsum in fibrous crystals is occasionally found, sometimes in amounts as great as 20%; it is an authigenic mineral.

Medium silt (0.031–0.011 mm): The composition is comparable to that of coarse silt, but here the percentage of glass shards is even greater (usually more than 60%). Montmorillonite, in aggregates, is also common and abundant. The amount of quartz is slightly increased; it can be as abundant as plagioclase, or more so. Orthoclase is scarce, but opal is present in small quantities.

Fine silt (0.011–0.0039 mm): Glass shards and montmorillonite aggregates make up more than three fourths of this fraction. Usually glass is greatly predominant over montmorillonite, but sometimes the reverse is true. The remaining minerals are, as usual, quartz, feldspars, rarely gypsum and opal.

Clay Fraction

In all the samples the clay fraction is composed of montmorillonite. This mineral appears in minute flakes; optical properties were studied with comparative ease on a film obtained by drying the suspension in a watch-glass. The individual flakes show parallel extinction, are length-slow and the optical figure is biaxial with small 2V. Measurements of refractive indexes gave the following results: α:1.512–1.25; γ:1.532–1.550.

X-ray diffraction data of the less-than-1μ fraction of six samples indicated the presence of a montmorillonite mineral as the main component; there were also small quantities of quartz, feldspar, and probably illite.

The remaining minerals of this fraction, as determined optically, are: volcanic glass, quartz, feldspars, chlorite, illite, kaolinite, etc. All these together very seldom constitute more than 10% of the fraction.

TABLE 3.—*Chemical analyses of Argentine Loess*

Location	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	CaO	MgO	K_2O	Na_2O	Ign. Loss	H_2O (120°)	Total
Alvear[1]	66.81	15.04	0.65	3.11	1.65	1.03	2.31	1.79	4.07	3.34	99.80
Alvear[2]	71.7	12.0	0.9	5.0	4.0	1.0	1.9	1.4	nd	0.5	100.5
Rosario[2]	65.5	15.6	1.1	6.7	2.1	1.4	1.6	3.1	nd	3.4	100.5
Baradero[3]	59.86	17.40	nd	4.80	3.08	1.71	1.71	1.97	3.22	6.04	99.79
La Plata[3]	62.70	15.00	nd	6.00	2.80	1.90	1.88	1.40†	3.38	4.32	99.38
Salto[3]	61.46	14.72	nd	4.91	2.46	1.85	1.56	1.74	4.56	7.11	100.37
Miramar[3]	57.16	17.28	nd	5.43	2.83	1.67	3.68†		3.60	8.35	100.00
La Plata[4]	63.34	16.13	0.74	5.10	3.46	1.64	5.79†		3.80	nd	100.00
La Plata[4]	63.34	17.72	0.80	4.95	2.35	1.84	4.33†		4.67	nd	100.00

† Determined by difference.
[1] Wright, F. E. and Fenner, C. N., 1912.
[2] Meigen, W. and Werling, P., 1915.
[3] Bade, P., 1920.
[4] Laboratorio de Química Analítica, LEMIT, 1955.

Chemical Composition

No chemical analyses were made for the present study. For purposes of comparison, some of the analyses found in the literature have been represented in table 3. The analyses of two samples from La Plata are new and were made at the Laboratorio de Ensayos de Materiales e Investigaciones Tecnológicas (LEMIT), of Buenos Aires Province.

$CaCO_3$ Content

Calcium carbonate is found in Pampean sediments as calcite in two forms: (1) as concretions, veins and hard-pan layers (caliche), which appear at different levels of the Formation. This is what locally is known as *tosca*, a name already registered by Darwin (1846); and (2) as comminuted particles distributed in the mass of the sediments.

The determination of calcium carbonate refers to the second type, as the first one is an epigenic deposit probably related to climatic variations. It is found that Argentine loessoid sediments have a lower content of calcite than the North American and European samples. This content varies from practically nothing up to a maximum of 8%. Most samples, however, usually contain less than 2% of calcium carbonate, and a few as much as 4%. This fact had been noted by previous workers (Scheidig, 1934).

SOURCE AND ORIGIN OF ARGENTINE LOESS

None of the important minerals composing the Argentine loess deposits are of local origin. The two ranges of the Province of Buenos Aires, the Sierrras de la Ventana (mostly quartzites) and the Sierras de Tandil (migmatites, gneisses, limestones, orthoquartzites), have not contributed in appreciable degree to the loess sedimentation. No data are available for similar deposits in the central and northern parts of Argentina, but there are indications that the Sierras de Córdoba have had some effect on the mineralogical composition.

The Argentine loessoid sediments are mainly formed by minerals of volcanic origin, especially those pertaining to andesitic and basaltic rocks. Characteristically metamorphic minerals are either scarce or absent. The really abundant constituents are plagioclases (labradorite and andesine, with smaller quantities of oligoclase and albite), which undoubtedly are derived from volcanic (and pyroclastic) rocks, as shown by their common zonal structure and the nature of the inclusions. On the other hand, quartz is rather scarce and, judging by the nature of its inclusions, mainly of igneous origin. Moreover, the constant occurrence of rounded fragments of volcanic rocks is further evidence in support of the petrographic nature of the predominant minerals. Some contribution has also been made by acid volcanic rocks, as is shown by the presence of felsitic groundmass fragments and altered orthoclase; it is supposed that these constituents are derived from rhyolites.

The heavy minerals are essentially represented by hornblende, pyroxenes (hypersthene and augite), and ores. It is difficult to establish which has been the parent rock of the amphiboles, but as they are common in modern volcanic ashes (Larsson, 1937) and in Tertiary and Quaternary tuffs (both amply distributed in Argentina), their derivation from these rocks is suspected. Hypersthene and augite (or diopside) are undoubtedly derived from vulcanites, since plutonic pyroxenic rocks are very scarce in Argentina. The same applies to the ores, which probably are also of volcanic origin,

In the finer grades the abundance of glass shards, either fresh or altered, is striking. This fact points to the important participation of pyroclastic rocks in the make-up of Argentine loessoid sediments. Admittedly, Argentina is a country in which volcanic activity has been very intense during the Tertiary and Quaternary, and as a result vast accumulations of pyroclastic materials cover a large part of its territory. But, in addition to this fact, it has to be remembered that apparently glass shards tend to concentrate in eolian sediments; this phenomenon may be due to their peculiar shape, which makes them more easily lifted by air currents, as pointed out by Russell (1936). The presence of montmorillonite as practically the only clay mineral adds still another proof of the importance of the pyroclastic contribution.

The establishment of the nature of the mineralogical constituents of Argentine

loess allows us to determine, in a rough way, the source areas of these minerals. Vast stretches of Patagonia and the Cordillera are covered with andesites and basalts, i.e., the rocks which may have provided most of the loess minerals. There are also in Patagonia large areas in which the predominant rocks are acid tuffs and rhyolites, fragments of which are found in the sand fraction of the Pampean Formation. Therefore, it is reasonably safe to assume that the source areas of the loess deposits were situated far to the west and southwest of the Pampas region.

The transport of the allochthonous material to the Pampas areas must have been effected by winds, since no important streams exist in the region. At present, the strongest and most frequent winds blow from the west and southwest. In 1932 a cordilleran volcano, the Quiza-pu, threw into the atmosphere an enormous volume of volcanic ashes, which were transported to the eastern part of Argentina by the westerly winds, as has been shown by the study of the dispersion of this pyroclastic material (Frenguelli, 1933; Larsson, 1937). It would seem, then, that the same atmospheric conditions prevailed during the Quaternary.

The wind-transported particles and grains which make up the loessoid deposits must have settled down slowly on the surface of the Pampas, where they were trapped by a thick grass cover; the existence of this vegetation is shown in the sediments by the numerous siliceous cells found in all the levels of the Pampean Formation. It is impossible to establish, on the basis of our present knowledge, whether or not the loess deposits were formed in the places where they are now found. However, since in certain levels of the Pampean Formation there are loess phenoclasts imbedded in a matrix of the same nature, it is to be assumed that superficial waters (rivulets, swamps, marshes, etc.) have played a certain role, either important or not, in the reworking and redistribution of the originally eolian sediment.

The majority of the loess minerals, with the exception of the potash feldspars, are completely fresh and devoid of alteration, in spite of the fact that many of them are relatively unstable. Most of the heavy minerals, such as hornblende, hypersthene, and augite, have endured intense wear and are, accordingly, very well rounded, but there are never signs of alteration. The conclusion to be drawn, then, is that they have undergone a long transport under arid conditions, and at the same time that conditions during and after deposition were not adequate for the alteration of the unstable species. The problem that still has to be solved is whether it was a hot or a cold arid climate that predominated during the formation of Argentine loess.

The lack of alteration of the loess minerals is even more striking in comparison with the abundant montmorillonized shards which make up the larger bulk of the fine silt and clay fractions. This fact must be explained without forgetting that, in the same sediments, there is always a considerable proportion of perfectly fresh glass shards and also that intercalations of unaltered volcanic ashes are frequent in the thickness of the Pampean Formation. The only seemingly acceptable explanation of this phenomenon is that the alteration of the shards into montmorillonite did not take place *in situ*. The incorporation of montmorillonite to the Pampean sediments, therefore, must have occurred as such, i.e., as material already altered, most probably derived from the volcanic tuffs so widely distributed in the western and southwestern parts of the Argentine territory. Montmorillonite, then, would be a "second hand" constituent, which has undergone at least two cycles of sedimentation: one, in Pliocene or early Pleistocene times, when the tuff deposits were formed and then altered; and another in the Pleistocene, when the formation of the loess deposits took place. As for the unaltered glass shards found in these sediments, they must be viewed as products of the volcanic explosions that intermittently punctuated the Quaternary and still occur.

Another point to be stressed is that no significant mineralogical or textural variations are found in the whole thickness of the Pampean Formation, i.e., during practically all the Pleistocene. It is then obvious that the same source areas, through similar

sedimentary processes, have contributed during all that time to the genesis of the loessoid sediments.

CONCLUSIONS

The results of this study allow us to draw some conclusions by comparison with North American and European loess.

1.—All loess samples seem to have similar granulometric composition; this applies equally to Argentine, North American, and European sediments, whether primary or secondary. Argentine loessoid sediments, like the others, are silts or sandy silts. The sand fraction is somewhat larger in Argentine samples, but it is mostly very fine sand, which lies practically on the border line separating it from coarse silt. The Mdϕ range is noticeably smaller in Argentine samples than in those of the northern hemisphere.

2.—The mineral assemblage composing the sand and coarse silt fractions of Argentine loesses is strikingly different from that found in North American or European samples. The abundance of plagioclases (mainly basic or intermediate), the scarcity of quartz, and the presence of a considerable amount of glass shards and fragments of volcanic rocks all point to a volcanic-pyroclastic origin. This conclusion is confirmed by the nature of the heavy minerals (hornblende, augite, hypersthene, and ores).

3.—The presence of abundant montmorillonite in Argentine loess is a point of similarity to the loess of North America, but of difference from the European. In our case, the origin of montmorillonite is evident, since in many samples it is possible to observe volcanic shards altered into that mineral. In this respect, the study of the Argentine loess has yielded clearer results than the North American samples. The observation of the montmorillonized shards, however, is only possible with direct microscopic investigation of the untreated samples, as dispersion of the samples destroys the aggregates of the clay mineral. This effect has to be taken into account in considering the granulometric composition, because the presence of these aggregates makes Argentine loess in the natural state coarser than what is shown by mechanical analysis.

It is evident, then, that loess is a sedimentary material which may vary considerably in mineralogical composition, but not so much in texture. The Argentine sediment, at least the one composing the Pampean Formation of Buenos Aires Province, is akin to what may be called a sedimentary tuff (Wentworth and Williams, 1932) or a tuffaceous silt (Hay, 1952), and constitutes another example of the importance of the pyroclastic contribution to terrestrial sedimentation, a question which has been recently stressed by Ross (1955). More research is needed to determine, in every particular case, the nature and origin of loess and its relation to climatic conditions (Bryan, 1945).

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FIG. 1, SEPTEMBER, 1957

SUGGESTED USE OF BED-THICKNESS MEASUREMENTS IN STRATIGRAPHIC DESCRIPTIONS[1]

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ABSTRACT

The description of stratification in sedimentary rocks in terms of the mean thickness and standard deviation of beds is suggested. The procedure is outlined and illustrated with an example from the Atoka Formation of Arkansas.

A problem commonly confronting the field geologist and stratigrapher is the description of stratification in measured sections of sedimentary rocks (McKee and Weir, 1953; Ingram, 1954; Gray, 1955; Bokman, 1956; Kelley, 1956). In many cases the data are presented in the form of a detailed table in which the thicknesses of successive beds appear in one column and the corresponding lithologic descriptions in another; or the stratification is described in subjective terms, such as "thin-bedded" or "massive" or "beds ranging in thickness from one inch to five feet." Both of these methods are somewhat less than satisfactory. Whereas the detailed table is merely a bulky enumeration of the raw data, the qualitative description is vague *per se* and is difficult to compare with other descriptions couched in similar terms. The writer submits that in many instances a quantitative summary would be equally useful and much more communicable than either a presentation of the complete data or a subjective statement of it. Such a summary can be made in simple statistical terms.

An example based on four groups of bed-thickness measurements from the Atoka Formation (Lower Pennsylvanian) of Arkansas (table 1) illustrates the application of this method to an actual field study. The stratification is described in terms of the mean and standard deviation, which summarize the important properties of the distributions. Both have obvious descriptive meaning, the one indicating the average thickness of beds in the section and the other the degree to which beds range in thickness about this mean, and may also have interpretive significance (Potter and Siever, 1955, p. 441–445; Kelley, 1956).

The first step in determining these parameters is to group the bed-thickness measurements on a logarithmic scale, such as the Wentworth scale for particle size. As shown in figure 1, this tends to normalize the distributions. The calculation of the mean and standard deviation may then be facilitated

TABLE 1.—*Summary of measured sections*

Sample	No. of beds	Bed thickness			
		Mean		Standard deviation	
		θ	Inches	θ	Inches
B1	96	1.67	3.18	0.49	1.40
B2	100	1.29	2.45	0.51	1.42
B3	100	0.10	1.07	0.24	1.18
JL	113	1.31	2.48	0.85	1.80
B2-JL	213	1.30	2.46	1.68	1.60

TABLE 2.—*Conversion of class limits from ξ to θ*

ξ (inches)	θ
128	7.0
90.5	6.5
64.0	6.0
45.3	5.5
32.0	5.0
22.6	4.5
16.0	4.0
11.3	3.5
8.0	3.0
5.7	2.5
4.0	2.0
2.8	1.5
2.0	1.0
1.4	0.5
1.0	0.0
0.7	−0.5
0.5	−1.0
0.4	−1.5
0.3	−2.0

[1] Manuscript received November 15, 1956.

by an adaptation of the Krumbein (1936) *phi* scale, in which the actual measurement and class limits are replaced by their logarithms to the base two. Since most bed-thickness measurements fall above one inch, positive instead of negative logarithms are desirable. Multiplying both sides of Krumbein's equation by minus one and replacing minus *phi* by *theta* gives

$$\theta = \log_2 \xi$$

where ξ is the bed thickness expressed in inches. Table 2 compares the class limits in inches with their *theta* equivalents.

In table 3 the calculation of the mean and standard deviation has been carried out for a sample of 213 beds. The equations for the mean and standard deviation, as well as the methods of determining them, are adapted from Dixon and Massey (1951, p. 17–22). The bed-thickness measurements are first grouped as in table 3, in which X_i refers to the *theta* midpoint of each group on the scale and f_i to the number of measurements in that group. The equation for the mean,

$$\bar{X} = \frac{\Sigma f_i X_i}{N}$$

states that all of the measurements are added and the sum divided by the total number of measurements. This is accomplished by dividing the total of column 5 in table 3 by the total of column 4. The equation for the variance of the distribution is

$$s^2 = \frac{\Sigma f_i X_i^2 - \frac{(\Sigma f_i X_i)^2}{N}}{N-1}$$

and may be computed as follows. From the total of column 6 subtract the squared total of column 5 divided by the total of column 4; then divide this difference by one less

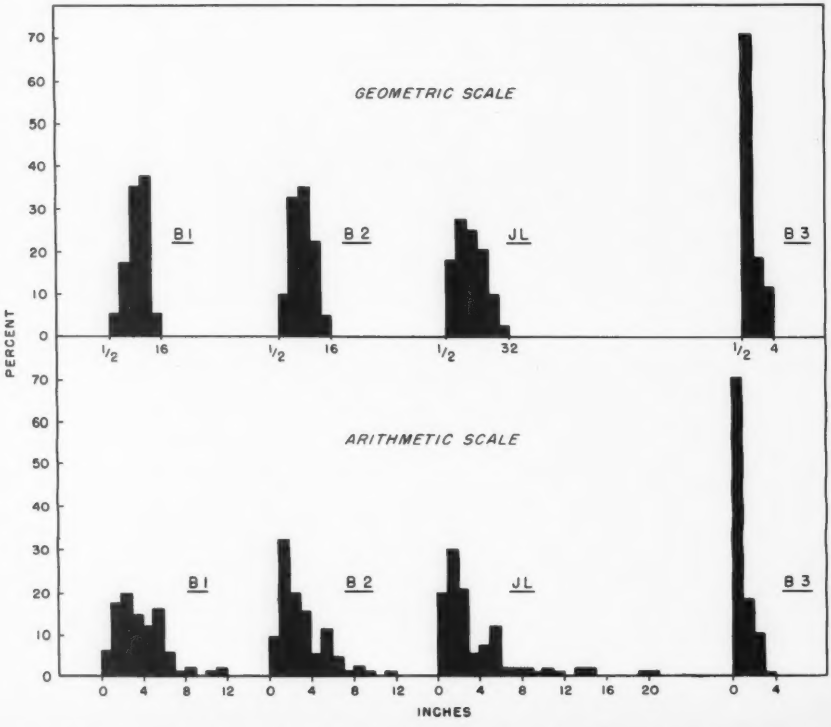

FIG. 1.—Histograms of bed-thickness measurements from the Atoka Formation. Comparison of corresponding histograms illustrates normalizing effect of plotting data on geometric scale.

TABLE 3.—*Arrangement of data for calculation of mean and standard deviation*

(1) Interval (inches)	(2) X_i	(3) X_i^2	(4) f_i	(5) f_iX_i	(6) $f_iX_i^2$
½–1	−0.5	0.25	29	−14.5	7.25
1–2	0.5	0.25	63	31.5	15.75
2–4	1.5	2.25	61	91.5	137.25
4–8	2.5	6.25	44	110.0	275.00
8–16	3.5	12.25	14	49.0	171.50
16–32	4.5	20.25	2	9.0	40.50
Totals			213	276.5	647.25

$$\overline{X} = 276.5/213.0 = 1.30\theta = 2.46 \text{ in.}$$

$$s^2 = \frac{647.25 - (276.5)^2/213.0}{212} = 1.36\theta = 2.57 \text{ in.}$$

$$s = \frac{1.36\theta}{2} = 0.68\theta = 1.60 \text{ in.}$$

than the total of column 4. One half of this number is the standard deviation.

The mean and standard deviation so calculated are in terms of *theta*. Conversion back to inches can be made approximately from table 2 or exactly from the "$(-\phi)$ mm." column of the *phi*-millimeter conversion table prepared by Page (1955).

The spread of the bed-thicknesses may be visualized by consulting table 4. The first column shows the spread in terms of standard deviation on either side of the mean and the second column indicates the percentage of the beds occurring within this interval. More accurate percentages for any desired interval may be obtained from standard statistical tables, such as tables 3 and 5 in Dixon and Massey (p. 305 and p. 307).

TABLE 4.—*Some approximate areas between $-z$ and $+z$ of the normal curve*

Standard deviations on either side of mean (z)	% of beds within intervals
0.3	25
0.7	50
1.0	68
1.2	75
1.6	90
2.0	95
2.6	99

These parameters may be used to summarize succinctly the stratification characteristics of an outcrop, core, or electric log and/or to compare it with other sections for which similar information is available. For example, samples B1, B2, and B3 were selected from different parts of the same outcrop which appeared to be distinct units and JL was drawn from an outcrop several miles distant. Comparison of the means (table 1) shows that the first three are, indeed, distinct groups but that B2 and JL are almost identical.

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FIGS. 1–2, SEPTEMBER, 1957

CLAY MINERALS FROM THE VENTURA BASIN, CALIFORNIA[1]

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ABSTRACT

Studies of clay fractions of sedimentary rocks from the Ventura Basin revealed a compositional difference between shales and sandstones. Montmorillonite or illite dominates in shale whereas penninite is most abundant in sandstone. Penninite increases in abundance with stratigraphic depth of the sandstone, its growth probably being favored by increased temperatures and pressures due to burial. Compositions of the shales also vary with stratigraphic position, but less regularly than the sandstones, possibly resulting from variation in proportions of allogenic clays supplied to the basin combined with different degrees of diagenetic alteration after deposition.

INTRODUCTION

Clay fractions of 125 samples of Pliocene sedimentary rocks from the western part of the Ventura Basin, California, were analyzed by X-ray diffraction methods. Clay minerals were identified and semi-quantitative estimates were made in order to learn of possible diagenetic changes. Outcrop and core samples of shale and sandstone were studied. Cores were taken from horizons recognized in wells of the Ventura Avenue Field, two miles north of Ventura, California (fig. 1). Surface samples were collected from stratigraphically equivalent horizons at the outcrop near Canada de Aliso, four miles north of the field. A total stratigraphic thickness of 8000 feet was sampled.

Each sample was disaggregated and the clay fraction was obtained by sedimentation. Many of the rocks are friable, but leaching of calcite cement by 2 N acetic acid was often required to free the clay. Disaggregated samples were dispersed using sodium hexametaphosphate, and all materials with equivalent diameters greater than four microns were removed by sedimentation. The remaining suspensions were agitated and allowed to settle onto glass slips to prepare oriented aggregates. The settling time was controlled so that the recovered aggregates had surface layers with particles near one micron in equivalent diameter. X-ray diffraction patterns were then made with a Norelco recording diffractometer using nickel-filtered, copper radiation. All X-rayed specimens were expanded with ethylene glycol. Additional diffraction patterns were obtained from more than half of the same samples in an air-dried state and after heat treatment. Semi-quantitative estimates of the percentages of the principal clay minerals were carried out using a modification of the method of Johns, Grim, and Bradley (1954).

[1] Manuscript received December 26, 1956.

DETERMINATION OF CLAY MINERALS

Identification

Montmorillonoid, mica, and chlorite minerals occur in the clay fractions of both shale and sandstone. The montmorillonoid mineral is identified on the basis of its variable 00*l* spacing under different degrees of hydration, and expansion of its basal spacing from 15.4 to 17 Å following ethylene glycol treatment of air-dry material. All measured dimensions of the 060 spacing lie near 1.50 Å, indicating a montmorillonoid of low iron content, a montmorillonite according to MacEwan (1951). A constituent with a 10 Å 00*l* spacing that is unaffected by standard heating and expansion techniques is designated as a mica, but the intensity of its 060 reflection is so weak that the spacing can not be determined with enough precision to establish whether the mineral is dioctahedral or trioctahedral. It is assumed that the mica is of the illite type.

A non-expanding mineral with a weak 14 Å and a strong 7 Å spacing is termed a chlorite. The 7 Å spacing is not appreciably affected by heating the specimen to 450° C, but heating to 550° C causes complete destruction of the 7 Å peak and an increase in intensity of the 14 Å peak. This test proves

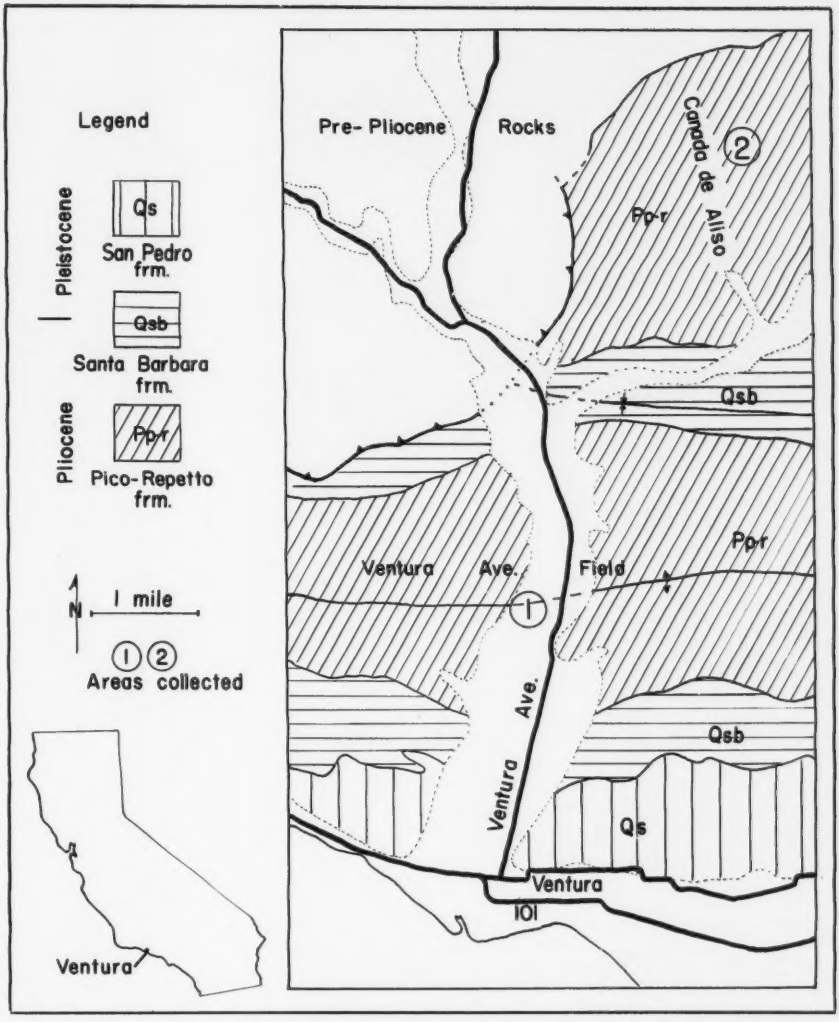

FIG. 1.—Index map showing the location and geologic setting of the Ventura Avenue Field and the sample localities.

the presence of chlorite, but not the absence of kaolinite, a mineral also having an 00*l* spacing at 7 Å which is destroyed at temperatures near 550° C. General reflections from unoriented powders do not correspond to those of well-crystallized kaolinite, and microscopic and differential thermal analyses fail to indicate its presence. For these reasons, the intensity of the 7 Å peak is considered to be due entirely to reflections from a chlorite mineral. The chlorite has an 060 spacing between 1.53 and 1.54 Å, indicating

a low iron content according to von Engelhardt (1942). Index of refraction measurements and birefringence estimations on coarse-grained material from the sandstones suggest that the mineral is penninite (Ny and Nz = 1.587 ± .003, Nx-Nz usually 0, ultra blue colors common).

Occasional randomly interlayered mica-montmorillonoid minerals were identified from low intensity reflections corresponding to basal spacings varying from 10 to 14 Å, but such interlayering is rare. Most of the interlayering noted is in clays from the deeper stratigraphic levels. Minerals other than clay which were recognized from their diffraction patterns include quartz, feldspar, and occasionally heulandite. X-ray amorphous material appears to be abundant in clays from the deeper horizons. Diffraction patterns of these materials show strong, low angle background radiation and weak reflections from all the clay components.

Semi-Quantitative Estimates

Estimates of clay mineral percentages were made by comparing the ratios of observed intensities of 00*l* reflections of particular components in natural mixtures with ratios of observed intensity in prepared mixtures containing known proportions of the same minerals. All intensity measurements were made on ethylene glycol saturated clay materials in which the glycol was introduced from the vapor state. Diffractometer traces of the first few orders of the basal reflections were made using a North American Phillips Company recording diffractometer. Relative intensity values were determined from the traces by measuring the areas under the peaks with a planimeter. Peaks used for comparison were the 17 Å reflection from montmorillonite, the 10 Å reflection from illite, and the 7 Å reflection from penninite. In this special case where kaolinite is lacking, the latter is valid. Relative intensity values of the reflections from each sample were related to the intensity value of the illite reflection in the form of ratios, penninite/illite and montmorillonite/illite. The ratios were then converted to weight percent values by comparison with experimentally determined curves. The curves used for comparison were determined from analyzed clay minerals having approximately the same iron content as the natural materials. This is necessary because the intensity of reflection for any given amount of material is strongly affected by the quantity and kind of ions present in the structure, particularly the number of iron ions. The prepared mixtures were X-rayed and two sets of curves of intensity ratios *versus* weight percent were plotted, one for the penninite to illite ratio and one for the montmorillonite to illite ratio. The observed intensity ratios determined from the natural mixtures were then compared with the ratios determined from the prepared mixtures. Weight percent values were assigned to the two components in each case. By setting the determined illite percentages equal and recalculating, percentages in terms of the three components were determined for each sample. Results of these estimates are presented graphically in figure 2 where mean values of mineral compositions over given stratigraphic ranges have been plotted against their stratigraphic position. In each case, the standard deviation of the mean and the vertical range of the samples considered are shown together with the number of determinations from which the mean was obtained.

DISTRIBUTION OF THE CLAY MINERALS

Sandstone *versus* Shale

Illite, montmorillonite, and penninite are present in both sandstones and shales of the Ventura Basin, but their percentage distribution in the two rock types is not the same. The most abundant clay mineral in the shale is montmorillonite, or sometimes illite, but penninite is usually scarce. In the sandstones, however, penninite generally predominates over illite and montmorillonite. Petrographic studies of the sandstones indicate that some of the penninite is allogenic, occurring in slightly rounded grains charged with opaque inclusions, but much of it is authigenic, for it fills pores and coats sand grains. Textural evidence, however, does not indicate whether the fine-grained penninite had an allogenic or authigenic origin. The different percentage distribution of clay minerals in sandstone as compared to shale is perhaps the most noteworthy of the observations recorded here. Being ultimately interstratified, these two rock types presum-

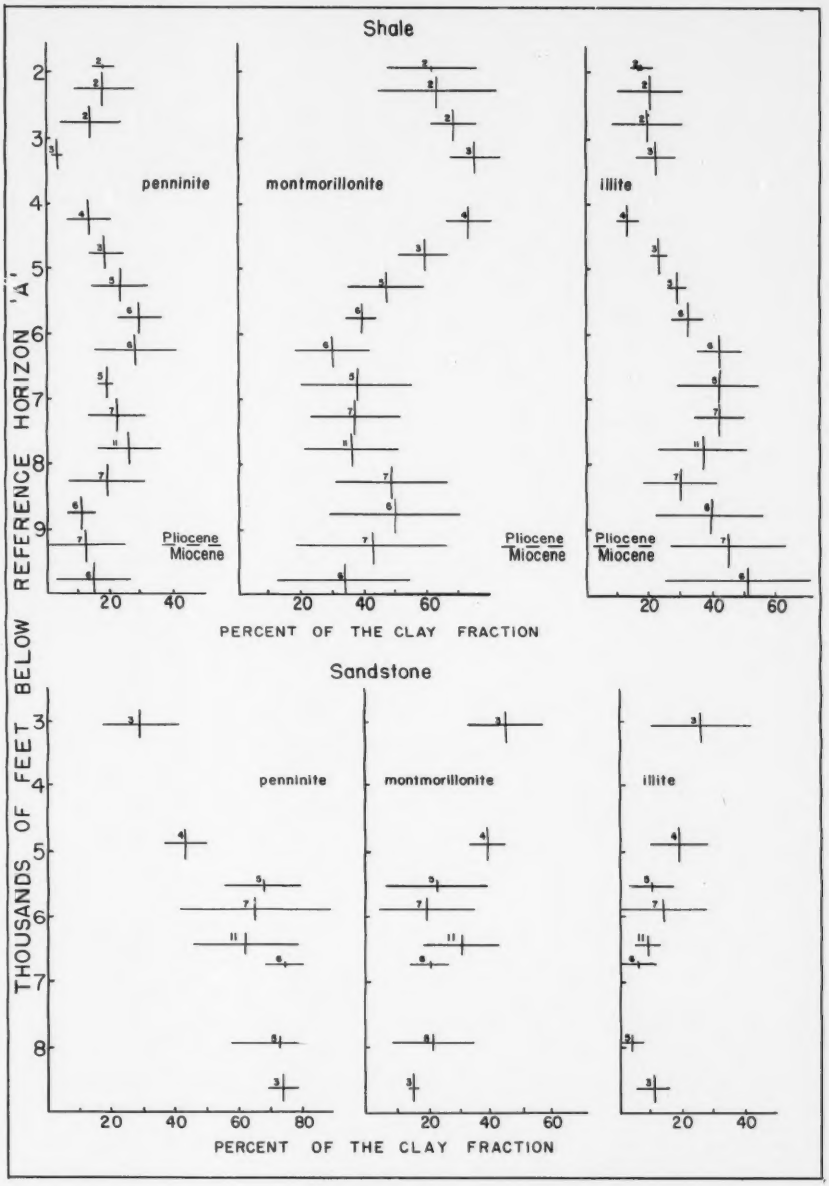

FIG. 2.—Variations of mineral composition of the clay fraction of shales and sandstones with changes in stratigraphic depth. The plotted points are mean values calculated from the number of determinations indicated in each case. The vertical line indicates the stratigraphic range of the samples and the horizontal line gives the standard deviation of the means.

ably had similar sources, and the mineral compositions were probably similar when the sediments arrived on the sea floor. The differences developed later. If the deposits were laid down by turbidity currents, as Natland and Kuenen (1951) indicate, the sands must have been deposited rapidly and soon covered by settling clouds of finer clay material which formed the shales. Clay trapped in the sand layers was identical to that in the covering shales. Textural relationships seen in thin section prove the diagenetic origin of at least some of the chlorite in the pores of the sandstone. Textural features further suggest that it formed after compaction of the sediment, for it is always associated with finely granular, late-formed calcite in tightly compacted rocks. Chlorite is never evident in the calcite cemented sandstones which preserve their original loose packing. Early diagenesis, therefore, was not important in the formation of chlorite in the Ventura Basin sediments. It developed after deposition, nourished by the connate waters which circulated most readily through the porous sandstones. It failed to develop in the shales because lower permeability reduced circulation of the waters.

Variations in Clay Minerals with Stratigraphic Level

The clay minerals also differ in relative amounts in the various stratigraphic levels. These differences are shown in figure 2 where the mean percentage values of the clay components of particular stratigraphic ranges have been plotted against their stratigraphic depths below a reference horizon. Standard deviations of each mean have also been plotted.

General trends in the mean values of mineral percentages *versus* stratigraphic depth can be recognized, even though the standard deviations of the means are often large. The large standard deviations are due mostly to real variations in the percentage compositions of the components. Experimental error also contributes toward the dispersion of points, but judging from the reproducibility of the values (within 10%), its contribution is small.

The variation in composition of the shales with stratigraphic depth is irregular. In general, montmorillonite is abundant in the upper horizons, but it decreases in amount to a minimum value near 25 percent at depths of 6000 to 7000 feet below the reference horizon. At greater depths no trend can be established because of the extreme scatter of the determined values. The penninite and illite trends are similar to one another. Both minerals are present in small amounts in the upper horizons, but increase slightly to a maximum at a depth of 6000 to 7000 feet.

Trends are also present in the sandstones. Penninite increases in amount downward from near 25 percent in the upper horizons to over 75 percent in the deepest levels. Simple statistical tests indicate that the trend toward increasing chlorite content suggested by the means of compositions in the stratigraphic range between 3000 and 5500 feet is not likely to be a chance variation, although admittedly the number of determinations upon which the trend is based are few. Trends in percentage variations cannot be recognized below 5500 feet, although the over-all aspect of the distribution is one of increasing chlorite content. Both montmorillonite and illite show a complementary decrease in amount downward, but montmorillonite has the greater change, from 50 percent in the upper levels to about 15 percent at depth.

That the clay mineral composition of the sediments varies with stratigraphic position is a second noteworthy conclusion in this study. In the sandstones the percentage of penninite increases in the lower horizons, and evidence has been presented which indicates a postdepositional origin for this material. The greater abundance of chlorite in the deeper sandstones may result from a longer period of time for authigenic growth, but more likely it reflects increased temperatures and pressures in the deeper horizons which favored penninite development in this particular chemical environment.

In the shales the variation of mineral composition with stratigraphic depth is more irregular and difficult to explain. The observed variations might result from changes in the supply of different allogenic clay particles during sedimentation, variation in early diagenesis rates due to changes in rates of deposition, or variations in diagenetic alteration after burial. Surely the

allogenic clay particles supplied to the Ventura Basin varied during early and middle Pliocene time, when the shales in question were laid down. Assemblages of heavy minerals, however, indicate no noticeable change in source rocks and no volcanic episodes. The source area consisted of igneous, metamorphic, and sedimentary rocks throughout. This does not, of course, eliminate the possibility that soil clays from the source areas might have varied in composition from time to time.

Late diagenetic changes undoubtedly took place in the shales as well as in the sandstones. Differences in the clay-mineral composition of the two rocks indicate that either late diagenesis has not been extensive in the shales or it has been entirely different from that in the sandstones. The latter seems to be more likely. Shale samples from successively deeper levels show a decrease in intensity of X-ray reflections from the clay minerals and a simultaneous increase in background radiation which is interpreted to indicate an alteration of clay minerals to X-ray amorphous material in the deeper levels. Preferential alteration of montmorillonite to X-ray amorphous material could account for the present distribution of clay minerals in the shales. Quantitative estimates of percentages of amorphous material were not attempted, however, and there is no evidence indicating the regularity of the suspected alteration.

The presence of mixed layer illite-montmorillonite minerals in the deeper horizons of the shales indicates another diagenetic change, probably in the sense montmorillonite to illite. Mixed layer minerals in these sediments are rare, however, suggesting that this alteration has not been extensive.

Effect of Surface Weathering

Clay mineral fractions from surface exposures in Canada de Aliso differ from those taken from the same horizons in the subsurface of the Ventura Avenue Field (fig. 1). Montmorillonite is everywhere dominant in the outcrop samples of both sandstone and shale, whereas illite and penninite occur only in trace amounts. By contrast in comparable subsurface samples, montmorillonite is sometimes the least abundant of the clay minerals. Such a difference in mineral assemblage over so short a distance suggests alteration of chlorite and illite to montmorillonite in the zone of weathering rather than original lateral variation in mineral assemblages.

ACKNOWLEDGMENTS

The writer wishes to express his gratitude to The Shell Oil Company for its contribution of well cores and structural information. Thanks are due also to The Standard Oil Company of California for its support of the research.

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FIGS. 1–4, SEPTEMBER, 1957

NOTE

A LIGHT-WEIGHT AUGER WITH QUICKLY DETACHABLE JOINTS[1]

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Hand augers are frequently the only convenient instruments for obtaining geological information in areas of unconsolidated rock. Brief references to the types available and their usefulness are made in such standard works on the sedimentary rocks as those by Milner (1952) and Krumbein and Pettijohn (1938).

The carpenter's auger, fitted to lengths of steel pipe has been popular for the last thirty years or so and (apart from various more elaborate and cumbersome devices produced commercially for post-hole sinking, and soil sampling) is the only type readily available to the geologist. These examples, while adequate for their purpose and undeniably robust, tend to be weighty and bulky, and often cause much waste of time in assembling and taking them apart.

With the prospect of making hundreds of shallow holes in the course of research on some Lower Cretaceous sands and clays, the writer set out to develop some improvements on the traditional form of apparatus.

The following requirements were felt to be desirable:

1.—The auger should be no undue encumbrance, from the point of view either of weight or size, to a single-handed worker wishing to cover a reasonable stretch of country in a day.
2.—The joints should be quickly made-up or broken in the field without the use of tools (spanners mean more weight to carry, and very often get lost in long grass).
3.—The bit should cut freely and be easy to remove from the hole. At the same time, it should retain a useful quantity of sample, which it should protect from contamination.
4.—The length of the rods should be standardized to facilitate rapid estimation of depth; their actual length should be such as to allow easy working.
5.—The design of the joints should be such that reversal of the direction of turning does not lead to unscrewing.
6.—It was also felt desirable, though not essential, to employ non-corrodable materials so far as possible.
7.—Finally, all the above requirements should be met in a design free from unnecessary complication, reasonably simple to construct, and sufficiently robust for field use.

These requirements were satisfactorily met in the design described here; it has been indicated in the description where field experience has shown that improvements could be made.

RODS AND JOINTS

It was felt that the rods usually employed were frequently too long. It is not easy, while turning, to apply weight to the handle of an auger if the latter is above waist height, nor is withdrawal easy under these circumstances. On the other hand, weight can be applied and turning easily carried out, from below waist height until the hands are within a few inches of the ground. A length of 2 ft. 6 ins. was accordingly adopted for all rods—a choice which experience confirmed to be a most comfortable one for a person of normal build.

After reviewing various types of joints which have been employed in the past, and considering a number of further possibilities, the design illustrated in figs. 1, 2, and 3 was arrived at.

The joint is a form of dovetail, the two parts being prevented from sliding out of place after connection by means of a sleeve,

[1] The appliance described here was developed in the course of research carried out at Imperial College, London, 1951–53, and was first described in a Ph.D. thesis entitled "The Petrography of the Shotover Ironsands" submitted to the University of London in April, 1954. Manuscript received November 15, 1956.

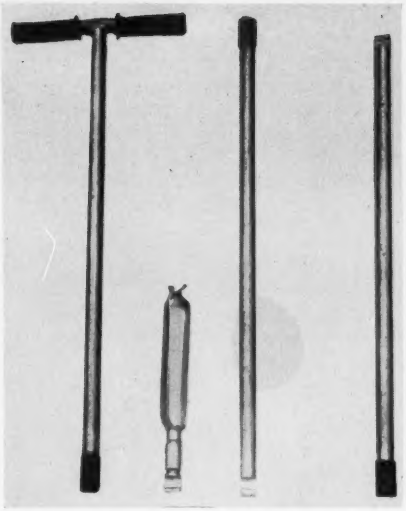

FIG. 1.—Auger rods and bit.

The proportions of the various parts of the joints are such as to give equal resistance to torsional stress across the weakest sections plus an adequate strength in tension to withstand the greatest pull likely to be exerted in practice.

In use, the sleeve (a) (fig. 2) is first moved along the rod so that part (b) of one rod can be slid sideways into (c) of the other. The sleeve is then moved back so as to cover up the joint, protecting it and preventing disengagement. The spring-loaded plunger (d) engages with the annular grooves turned inside the sleeve at each end, preventing it from being withdrawn completely from the rod, and locking it (against accidental movement) in either of its two extreme positions. The sequence of events is shown in fig. 3.

Making or breaking a joint of this kind takes only a matter of seconds. The joint, when made up, is stronger than it might appear at first glance and if well made, quite rigid.

In the example made by the writer, tube and joints were of aluminum alloy in order to save weight, prevent corrosion, and facilitate machining. At the time, the only tube available had a wall thickness of only $\frac{1}{16}$ inch. The use of this thin material led to failure at the weld between the tube and the joint on two occasions of abnormal stress in service. Apart from this failure, the joints themselves and the tubes forming the rods were found to be quite strong enough to stand up to hard work, although slightly more care is required than when using $\frac{1}{2}$ inch gas pipe. When the photographs were taken, the equipment had penetrated a cumulative total of well over one thousand feet of sediments.

The use of aluminum alloy did, however, give rise to one other trouble. There was occasionally a tendency for the sleeves to stick as a result of particles of grit becoming embedded in the relatively soft metal. This difficulty can be almost entirely eliminated if care is taken to keep the joints wiped

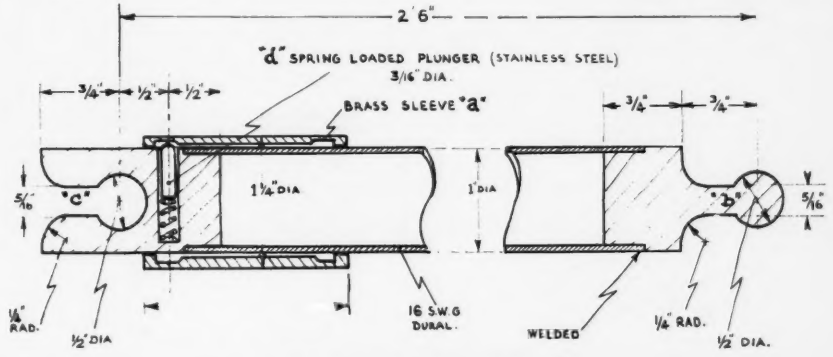

FIG. 2.—Design of auger joints.

FIG. 3.

clean, but a more complete solution might be to employ a harder material for joints and sleeves. Case hardened mild steel or stainless steel would be ideal from this point of view, but plain mild steel would probably be adequate. If this were done it would still be possible to use light-alloy tubing for the rest of the rods if desired—it is not widely appreciated that techniques exist for welding certain light alloys to certain steels—in which case an increase of wall thickness to about $\frac{3}{32}$ inch or even $\frac{1}{8}$ inch might be an advantage.

The proportions of the handle also received some thought. Very often this item is made too long, and the annoyance of the ends catching in clothing at each revolution needs no elaboration to anyone who has used the usual equipment—particularly in wet weather. A width of one foot was found to provide all the leverage necessary for boring a 2 inch diameter hole with the type of bit described below. In the example made by the writer the handle was simply a length of tube welded "T" wise to the end of one

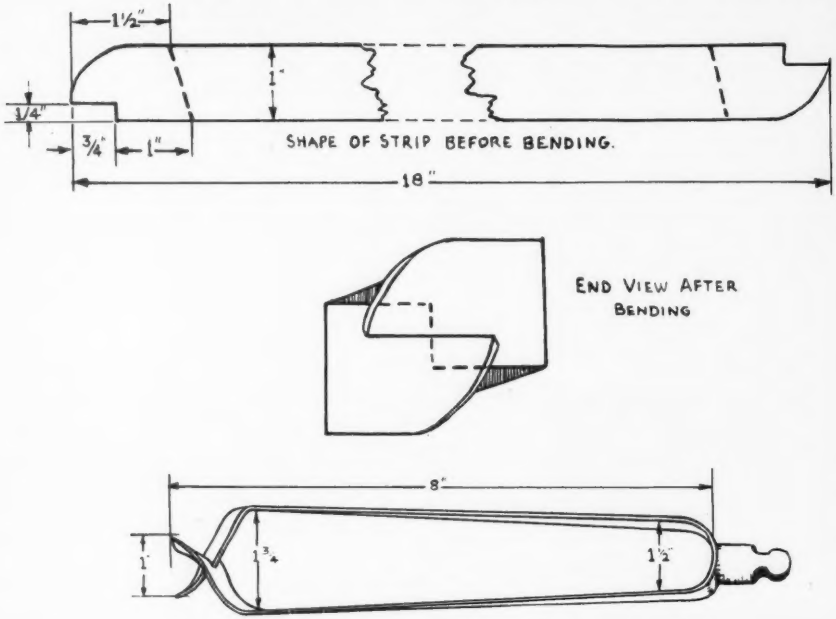

FIG. 4.—Sketch showing construction of auger bit. Not to scale.

of the rods. A detatchable handle could alternatively be fitted, using a standard joint of the type just described. The diameter of the handle should not be less than one inch for comfort in use, and this part can conveniently be fitted with standard motorcycle type rubber hand grips.

Three lengths of rod of the type described were made, giving a total length of 8 ft which is usually adequate for confirming boundaries and obtaining relatively unweathered samples in temperate climates.

One rod, complete with joints and sleeve, weighed just under 400 gm, compared with a weight of 965 gm for the same length of ½ inch steel gas pipe with one screwed sleeve.

THE BIT

No particular originality is claimed for the design of the bit; rather, it represents the logical development of an old type, in the directions of simplicity and cheapness of construction, and ease in use.

Unlike the spiral carpenters auger, this bit can be run to its full depth in tough, plastic clays without sticking. This is due to the fact that it is so proportioned as to provide clearance with the hole in all directions from the cutting lips. Sand (unless unusually dry or wet), clay, and soft shale, are handled with ease, and a relatively uncontaminated sample of about 500 gm obtained. Gravel, if it has a fairly high proportion of fines and few angular pieces larger than ¾ inch, can also be penetrated, though with less facility.

Figure 4 shows the proportions of a bit to cut a hole 2 inches in diameter. It is made from a single strip of 1 inch×3/32 inch mild steel about 18 inches long. The ends are filed to the shape shown and then bent along the dotted lines so as to make an angle of about 120° with the straight portion. The strip is then bent about its midpoint to a radius of about ¾ inch, until the lips come together, when they may be secured by spot welding where they overlap. A slight twist gives the lips a more correct spiral form and assists penetration, while relieving the cutting edges with a file helps the cutting action. The boss carrying the joint is welded on at the bend, after which the lower half of the tool can be case hardened.

CONCLUSION

The auger described here proved itself in field conditions to be equally useful for rapid confirmation of lithological boundaries under soil cover and for the collection of samples for laboratory study of an area of unconsolidated sediments. The special form of joint devised was found to be basically sound and durable, while the care taken in choosing the proportions of such apparently trivial details as the handle, and other details of design, was found to be amply repaid by ease and convenience of operation. By eliminating many of the annoyances and delays inherent in the use of the older types of equipment, the writer feels that his design can contribute towards the collection of better field data, for there can be little doubt that the fewer material handicaps the geologist encounters in the field from his equipment, the more attention can he give to the purely geological problems involved in his work.

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FIG. 1, SEPTEMBER, 1957

DISCUSSION

REPLY TO THE "DISCUSSION" OF A FABRIC AND PETROLOGIC STUDY OF THE PLEASANTVIEW SANDSTONE[1]

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In the "Discussion" presented in the June, 1957, issue of the Journal of Sedimentary Petrology (v. 27, no. 2, p. 198–201) John E. Sanders has taken exception with several statements made by Rusnak in a paper entitled "A fabric and petrologic study of the Pleasantview Sandstone" (Rusnak, 1957). From this discussion it would appear that there have been some misunderstandings of the text and illustrations presented by Rusnak which require further clarification. Sanders objects in particular to (1) the comparison which was made of the Pleasantview depositional conditions with those prevailing in the Dutch Wadden Sea, and (2) Rusnak's use of the term "flow rolls." In order to reply to Sanders' first objection, it is necessary to restate the writer's reasons for suggesting a comparison with the Dutch Wadden Sea.

The Pleasantview Sandstone is the basal member of the Summum (Pennsylvanian) Cyclothem of western Illinois. This sandstone occupies deeply cut channels and can be conveniently described as two-phased; the thin (6–10 feet) non-channel phase, and the thick (up to 100 feet) channel phase. The non-channel phase is thin-bedded, wave-ripple marked, calcareous, slightly carbonaceous, and in apparent conformity with the underlying silty shales. The channel phase is massive to thin-bedded, cross-bedded, shaly, very carbonaceous, and contains a locally derived basal conglomerate and is uncomformable with the underlying channel-cut beds. A characteristic feature of the channel sandstone is the initial dip of the bedding, which dips towards the channel axis along the channel walls and assumes an essentially horizontal attitude in the central part of the channel. These relationships, along with the evidences of slump structures and scour marks within the channel, are considered by the writer as indications that the Pleasantview sandstone was deposited when both the channel and non-channel areas were submerged. The non-channel areas are thought to represent continuous gradational change of sedimentation from silty shale to sandstone and the channels represent areas of contemporaneous erosion and later filling.

Measurements made of the highly variable cross-bedding, indicated that the direction of sand transport (within the single channel studied) was up channel and contrary to the channel gradient. In addition, the studies of sand-grain orientation (in uncross-bedded units) indicated that some sand transport occurred across the channel axis. The sediment transport directions would seem, therefore, to indicate that the process of channel filling was largely independent of the process or processes which cut the channel.

An evaluation of the specific features in the Pleasantview Sandstone, in terms of a possible present day counterpart, lead the writer to believe that these features had many similarities with those reported from the tidal flats and tidal channels of the modern Wadden Sea off the northern coast of the Netherlands. Table 1 shows these similarities and includes some remarks on this comparison. The properties of the Wadden Sea deposits presented in this table are based on the published descriptions given by van Straaten (1951, 1952, 1954a, 1954b).

This comparison indicates that the apparent environment of deposition of the Pleasantview Sandstone is not greatly dissimilar from that of the Dutch Wadden Sea deposits. The greatest discrepancy in this comparison is the absence of marine fossils in the Pleasantview. Nonetheless, the writer suggests that the sedimentological

[1] Contribution from the Scripps Institution of Oceanography, New Series 944. Manuscript received June 26, 1957.

TABLE 1.—*Comparison of environment of deposition of Pleasantview Sandstone with the Dutch Wadden Sea*

Similarities	Remarks
1. Channels similar in shape and size	Sedimentation essentially continuous outside the channeled areas
2. Presence of locally derived mud-pebble basal conglomerate in the channels	These basal conglomerates are considered as lag deposits derived from the eroded walls of channel-cut sediments. The Wadden Sea conglomerates commonly consist of shell beds. No shells or evidence of organisms noted in the Pleasantview at the locality studied
3. General lithology and grain size	
4. Irregular cross-bedding	In the Wadden Sea the cross-bedding may be directed up channel in areas subjected mainly to flood tides and may be directed down channel in areas dominated by ebb tides
5. Presence of slump structures	
6. Wave and current ripple marks	Ripple marks in the Pleasantview are found in the non-channel areas, but are rare in the channels. They are common to both areas in the Wadden Sea
7. Great abundance of drifted plant debris in channels	
8. Irregular stratification	
9.	Although mud cracks are common to the Wadden Sea area; no mud cracks were noted in the Pleasantview
10. Inclined initial bedding along channel walls	The mechanism of forming the inclined bedding in the Wadden Sea channels is not the same as that of the Pleasantview

properties of the Pleasantview Sandstone can be explained best with an environment similar to the Wadden Sea tidal flats. Under such conditions the channel is first cut by tidal current action and then filled by a combination of lessened tidal currents sweeping sand up the channel and longshore current drifting of sand across the channel.

Sanders questions such a comparison with the Dutch Wadden Sea mainly on the basis of the structural relations of the channel-filled to channel-cut sediments. Figure 1 illustrates these relations in the Pleasantview and is a modification of the cross-section presented by Ekblaw (1931, fig. 4, p. 386). The figure used by Sanders (1957, fig. 1b) to illustrate this structure is a misapprehension of the description given by the writer (Rusnak, 1957, p. 43). However, the important thing to note is that the sandstone beds flanking the channel walls are inclined towards the channel axis, while near the center of the channel the bedding is essentially horizontal. This channel structure is similar to the smaller scale channel structures observed in the Wadden Sea by van Straaten (1951, 1952, 1954a, 1954b). However, although there is this structural similarity, the mechanism responsible for the inclined bedding along the Pleasantview channel is not comparable with that of the Wadden Sea channels. In this regard Sanders argues correctly against the application of the term "lateral sedimentation" to the inclined bedding along the Pleasantview channel. Nonetheless, without due regard to the numerous similarities between the Pleasantview Sandstone and the Wadden Sea deposits Sanders rejected the environmental comparison on the basis that the inclined beds of the Pleasantview are not due to "lateral sedimentation."[2]

[2] Sanders (1957, p. 198) implied that this comparison of environments by Rusnak (1957) was made solely on the basis of similarities in initial inclined bedding along channel walls. He states, "In view of this relationship, Rusnak compares the conditions of the Pennsylvanian Pleasantview

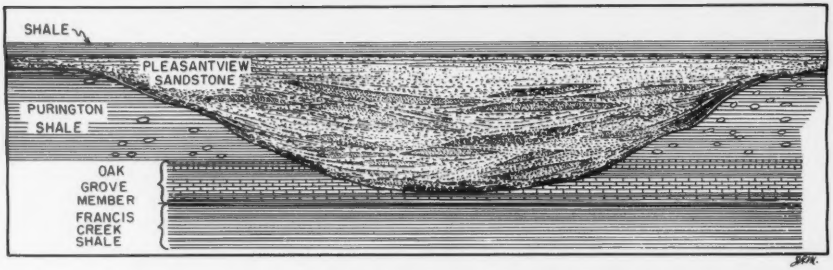

FIG. 1.—Generalized cross section of the Pleasantview channel exposed in outcrops along Coal Creek, Fulton County, Illinois (after the figure and interpretation by Ekblaw, 1931, fig. 4, p. 386). Depth of the channel in the central portion is about 50 to 55 feet. Vertical exaggeration ×33.

The term "lateral sedimentation" was used and defined first by van Straaten (1951, p. 236–238, 243; 1952, p. 500–502; 1954a, p. 26–27; 1954b, p. 7–8) for the specific mechanism associated with the development of inclined bedding in the Wadden Sea channel deposits. A detailed discussion of this mechanism and its resulting bedding need not be repeated here because this has already been elaborated upon by Sanders (1957). It is sufficient to say that the term can be applied only to inclined strata which are deposited laterally on point bars, such as those of tidal meanders, as Sanders rightly points out.

Lateral deposition could be only of minor importance in forming inclined bedding in the Pleasantview channel deposits. The great depth of this channel and the observed sand-grain orientation direction, which suggests deposition across the channel axis, would preclude such a process. The writer concluded, therefore, that the dominant mechanism which developed this inclined bedding consisted of a sweeping in of sediment from the surrounding area (Rusnak, 1957, p. 43, 52; for a generalization of other channel sandstones, see also Wanless and Shepard, 1936, p. 1202; and Weller, 1956, p. 29). It may well be, however, that some of the small-scale inclined bedding, associated with apparent meandering within the channel, is due to "lateral sedimentation."

Sanders further pursues his argument against a comparison with the Wadden Sea channels by suggesting four possible combinations of mechanisms to explain the origin and subsequent filling of the Pleasantview channels. These suggestions are:

"1) tidal channels—littoral marine filling, perhapes indicating subsidence but no particular changes in geography; filling sediment of marine origin. 2) tidal channels—fluvial filling, indicating marine regression between cutting and filling; filling sediment non-marine. 3) fluvial channels—fluvial filling, indicating a change in stream regime from cutting to filling; filling sediment non-marine. 4) fluvial channels—littoral marine filling, indicating marine transgression; filling sediment should contain marine fossils."

Of these possible combinations, Sanders concludes, "As the basal channel sand of the cyclothem generally passes upward into coal deposits, the filling sediment is doubtless non-marine in most instances, though the channel could be of either tidal (alternative 2) or fluvial (alternative 3) origin." There are, however, several objections to these conclusions which can be resolved only by the first suggestion. Both suggestions 2 and 3 are based on the assumption that the channel deposits are of fluvial origin. This assumption arises from the fact that generally neither the channel sandstones[3] nor

Sandstone with those prevailing in the modern Wadden Sea off the northern coast of The Netherlands." That such an implication is in error, is demonstrated by the itemized list of comparable features from both areas listed under the heading "Comparison with the Dutch Wadden Sea" (Rusnak, 1957, p. 51) and amplified in the present paper (table 1).

[3] Ekblaw (1930) reports marine fossils from the Pleasantview sandstone in the area farther south than that which was studied by Rusnak (1957). This occurrence of marine fossils in the Pleasantview may represent another facies of the sandstone which was more favorable to marine organisms.

the overlying shales contain marine fossils. However, this assumption is not necessarily correct and is problematical. All investigators of the Pennsylvanian are not agreed that this assumption is valid (Wanless, 1950, p. 25). Indeed, Wanless (1952, p. 150–152) has suggested that the stratigraphically lower Seville Cyclothem channel (containing marine limestone) had been cut by streams or tidal currents and filled under deltaic of littoral conditions. Wanless has also suggested (with reservations) that the conditions of the Pleasantview were comparable (1952, p. 152). One should not, therefore, apply generalizations indiscriminately.

Perhaps the most serious objection to fluvial deposition of the Pleasantview lies in the observations of cross-bedding. These observations, although limited, indicate that the direction of sediment transport was in part up channel and contrary to the erosional gradient of the channel. It is difficult to conceive of a method whereby fluvial processes could form cross-bedding directed contrary to the bottom gradient since the water transport would have to follow the gradient. In the writer's opinion, this evidence alone precludes the possibility of fluvial channel filling. This apparent contradiction can be explained simply, however, by having tidal currents sweep sediments up the channel, as has been demonstrated by van Straaten (1950). Moreover, this mechanism may account for the variability of cross-bedding dip directions (see remarks in table 1).

The writer, furthermore, concluded that it is necessary to have two mechanisms operative during the channel filling stage in order to explain the initial dip along channel walls and the apparent discrepancy of grain-orientation direction from cross-bedding direction. For this purpose he proposed to explain the inclined bedding and grain-orientation direction as resulting from drifting of sand across the channel axis; while, the more nearly horizontal, massive and cross-bedded sandstone in the central section of the channel resulted from strong tidal currents sweeping sediment up the channel (Rusnak, 1957, p. 50, 52).

It was not the intent of the writer to solve regional problems of cyclothem channel cutting and filling with the very limited local study. He did try, however, to explain the observed phenomena from this single channel as best fitted the evidence. Whether the proposed mechanism of channel cutting and filling is valid and can be applied to more cyclothem channel systems remains to be seen. For the present, the writer rejects Sanders' suggestion of a fluvial origin for the Pleasantview channel filling, in the area investigated, and proposes instead to adhere to the original suggestion of a tidal flat environment similar to the Dutch Wadden Sea (Rusnak, 1957, p. 52).

Sanders' remarks concerning the use of the term "flow rolls" in Rusnak's paper (1957, p. 43) are in part correct. That is, not all "flow rolls" can be attributed to subaqueous scour. If the writer seemed to imply that Rich (1950, p. 727) attributed all "flow rolls" to scour, the implication was completely unintentional. Rich clearly stated that these markings can develop also by load flowage of unconsolidated sediments into underlying soft beds. The writer is fully aware of the distinction which Rich and Kuenen (1953, p. 23–25; and more recently 1957, p. 231–258) made of the causes of these markings. However, he believes that the marks described in the Pleasantview as "flow rolls" (and associated with channel walls) are scour casts rather than load casts. These features were examined in great detail in an effort to distinguish their mode of origin. In both the "groove casts" and the "flow rolls" there was no evidence to indicate structural deformation of the underlying surface. Furthermore, the casts were cut through on a diamond saw and examined for internal structure. No internal structure was observed and, hence, the writer concluded that these features were of the type described by Rich as scour markings.

Both load flowage and scour features are present in the Pleasantview as is obvious from the figures presented by Rusnak (1957). Where load flowage was clearly indicated by internal structure the features were described as such.

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REVIEW

Illustrations of the Huttonian Theory of the Earth, by John Playfair. Pp. xx+528; 5½×8½ in., cloth. Originally published in Edinburgh, Scotland, in 1802; facsimile reprint, with introduction by George W. White, published in 1956 by University of Illinois Press, Urbana, Ill. Price $4.50.

Playfair's widely-quoted geologic classic, which played so significant a role in the history of modern geology, has long been out of print, is lacking even in many geologic libraries, and has been obtainable only as a collector's item at the prices which that usually implies. The publication of a reprint at the modest price specified is therefore a most welcome event. For teachers and students a long-felt need is filled, and for all geologists with a philosophical turn of mind a source-book of great value is made readily available. The book is a must for every professional library.

As the title implies, Playfair's work is devoted to expounding the geological concepts of James Hutton, with whom modern geological thinking began. Hutton's voluminous and somewhat ponderous writings are reduced to their essential concepts, and presented in lucid, logical, and readable prose—a fact which played a significant part in ultimately gaining their acceptance. Roughly the first quarter of the book is given over to a systematic presentation of the Huttonian theory, and the remaining part contains explanatory notes, elaborating on numerous points, presenting additional evidence, and refuting the claims made by critics and by proponents of competing schools of thought. The systematic part is divided into three sections: (1) Phenomena peculiar to stratified bodies, (2) Phenomena peculiar to unstratified bodies, and (3) Phenomena common to stratified and unstratified bodies.

The first section and accompanying notes, dealing with stratified bodies, should be of particular interest to readers of this journal. Here we find a clear realization that all "present strata are composed of the remains of more ancient rocks," a viewpoint in direct conflict with the then-prevalent Neptunist school of Werner. The origins of calcareous rocks, siliceous rocks, argillaceous rocks, and coal are considered in turn. The one major departure from modern thinking has to do with the "consolidation of the strata," which Hutton attributed primarily to heat. In discussing the "position of the strata," departures from original level of deposition, as seen in deformed strata, are analyzed, the effects of original dip are noted, and the significance of angular unconformities is recognized. It is shown that changes in sea level alone are inadequate to explain these phenomena, as had been supposed by others, and that "a force directed from below upwards" is required; this is attributed to "the expansive power of heat."

The second section, concerned with unstratified bodies, is of particular interest to students of igneous rocks. The third section and its notes, devoted to phenomena associated with both stratified and unstratified rocks, takes up weathering, erosion, transportation and "the great geological cycle, by which the waste and reproduction of entire continents is circumscribed." Here we discover the roots of modern concepts in both geomorphology and sedimentation, developed by rigorous reasoning on the basis of the present as a key to the past. Indeed it is almost surprising to note that so many ideas which we now take for granted then required support from detailed argument. We cannot but be impressed by the extent to which so many things which now seem inescapably obvious became so only through a long process of selection and rejection, testing and proving. Despite their unassailable logic, the views of Hutton and Playfair made headway only gradually against the orthodoxy of their day, aided in time by Sir Charles Lyell's brilliant advocacy, and by the impact of a constantly growing body of field observation from two continents.

It may be remarked that not all of the ideas championed by Hutton were necessarily original with him. Various elements of his system were anticipated, piecemeal, by Steno, Lehmann, Fuchsel, Guettard, Desmarest, and others. It was Hutton, however, who first incorporated these and other ele-

ments into one coherent system with universal applicability, solidly based at every point on logical reasoning from specific field facts.

Playfair's book, although outstanding for its early exposition of the basic tenets of modern physical geology, is equally noteworthy for its mode of approach. It may, in fact, be regarded as a primer on the application of the scientific method to geological investigation. At a time of casual observation, unbridled speculation, and boundless credulity, it was distinguished by its judicious tone and its continuous appeal to observational data as a basis for inductive reasoning. In contrast to the theories of various contemporaries, it dealt only with systematic operation of uniform and natural forces, susceptible to observation, and not with accidental or supernatural causes outside the realm of observation and analysis. Playfair realized fully the limitations of the data available to him, and the resulting imperfections of his system, stating that "it cannot be doubted that the great principles of it, though established on an immoveable basis, must yet undergo many modifications, requiring to be limited, in one place, or to be extended, in another." (p. 139) "Again, theories that have a rational object, though they be false or imperfect in their principles, are for the most part approximations to the truth, suited to the information at the time when they were proposed. They are steps, therefore, in the advancement of knowledge, and are terms of a series that must end when the real laws of nature are discovered." (p. 512) Playfair was firmly confident that the application of his philosophy of observation and critical reasoning would in due time make any corrections that might be required in his theory.

Adequate appraisal of current geological thinking requires that it be viewed in the perspective of its historical development. Playfair's volume represents a major turning point in that development.

H. T. U. Smith
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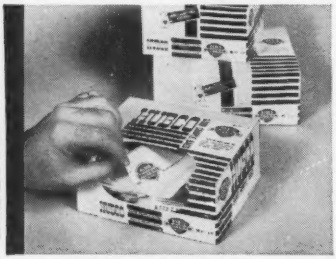

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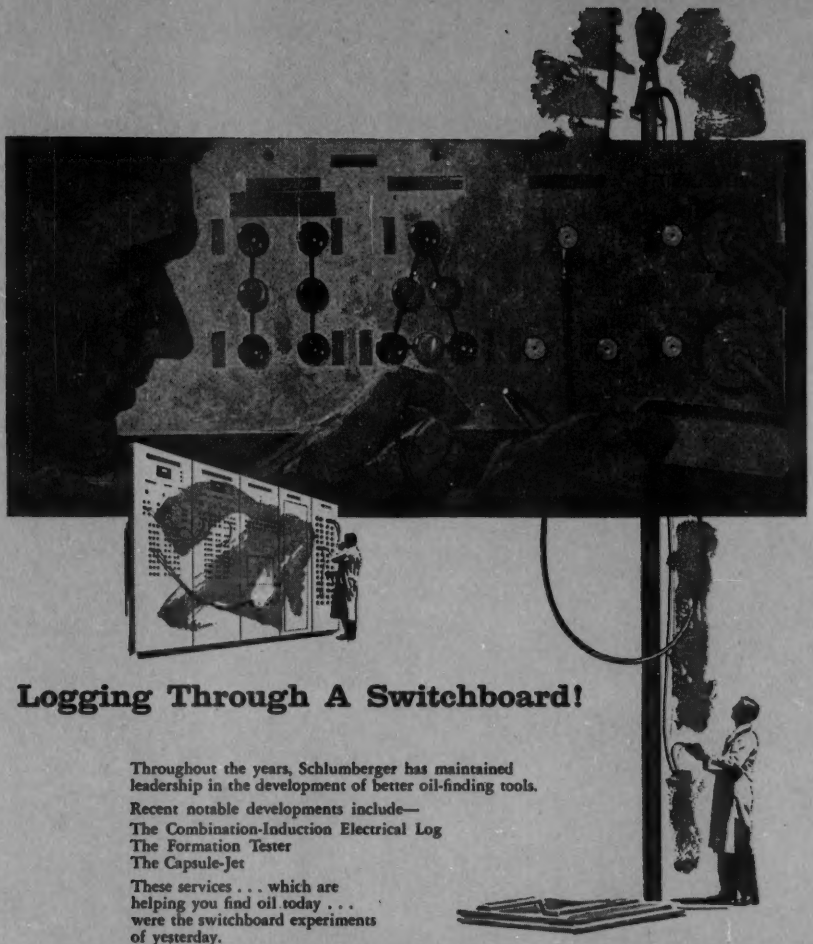

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